ELEMENTARY

METEOROLOGY

BY

WILLIAM MORRIS DAVIS

PROFESSOR OF PHYSICAL GEOGRAPHY IN HARVARD COLLEGE

GINN AND COMPANY

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ELEMENTARY METEOROLOGY.

CHAPTER I.

THE GENERAL RELATIONS OF THE ATMOSPHERE.

1. The subject of meteorology. We dwell on the surface of the land; we sail across the surface of the sea; but we live at the bottom of the atmosphere. Its changes pass over our heads; its continual fluctuations control our labors. Whether our occupation is indoor or out, on land or at sea, we are all more or less influenced by changes from the clear sunshine of blue skies to the dark shadows under clouds; from the dusty weather of droughts to the rains of passing storms; from the enervating southerly winds to the bracing currents from the north. Few persons fail to raise some question now and then concerning the causes and processes of these changes; some inquire more earnestly, desiring to inform themselves carefully on the subject. No school study suggests more frequent questions from scholars, or allows more educative replies from teachers than meteorology, the science of the atmosphere.

It is the author's intention in preparing this book to place before both readers and students an outline of what is now known in the domain of meteorology; to give a condensed account of the present condition of the science, without too much technical language or argumentative demonstration. All available sources of information have been drawn upon in the effort to make the various chapters represent the position of the modern meteorologist.

2. The plan of this book may be concisely stated. The origin and uses of the atmosphere are first considered, with its extent and arrangement around the earth. Then, as the winds depend on differences of temperature over the world, the control of the temperature of the atmosphere by the sun is discussed, and the actual distribution and variations of temperature are examined. Next follows an account of the motions of the atmosphere in the general and local winds; in the steady trades of the torrid zone and in the variable westerly winds of our latitudes. The moisture of the atmosphere is then studied with regard to its origin, its distribution and its condensation into dew, frost and clouds. After this, we are led to the discussion of those more or less frequent disturbances which we place together under the name of storms; some of them being large, like the great cyclones or areas of low pressure on our weather maps; some of them very small, like the destructive tornadoes. The effect of these storms and of other processes in the precipita-

tion of moisture as rain, snow and hail is next considered. Closing chapters are then given to the succession of atmospheric phenomena that ordinarily follow one another, on which our local variations of weather depend, together with some account of weather prediction; and another on the recurrent average conditions that we may expect in successive seasons, repeated year after year, which we call climate.

3. Meteorology as a branch of physics. All the conditions and phenomena of the atmosphere are illustrations of the principles of physics. The properties of gases and vapors, and the laws of heat and motion are here exemplified on a great scale, vastly larger than that usually considered in laboratory experiments; but the difference of scale does not in any way affect the application of physical laws.

It is therefore essential that the student should have at least a fair elementary knowledge of physics, gained if possible from laboratory experiments as well as from the study of text-books, before entering on the subject of meteorology. If any such terms as the following are not precisely understood, they should be carefully studied again in a good book on physics as they are encountered in these pages: mass, volume, density; inertia, force, velocity, rotation, centrifugal force; gravitation, gravity, weight; atom, molecule; solid, liquid, gas; expansion, heat, temperature, specific heat, latent heat.

Origin of the Atmosphere.

4. Relation of the earth to the other planets. The atmosphere, chiefly a mixture of nitrogen and oxygen, is thought to be a thin remainder of a once much larger volume of denser gases and vapors. Our understanding of this comes best by looking into the early history of the earth and the other planets that accompany the sun. All these planets are nearly spherical bodies, rotating as far as known from west to east and moving around their orbits in the same direction. Most of them are accompanied by one or more satellites, revolving again in the same direction. The sun turns on its axis in the same way as the planets revolve around it.

The resemblances among these bodies are indeed so numerous and so striking that it has come to be generally believed that the matter of which they are composed was once scattered thinly through an enormous space, making a vast cloud or *nebula*, similar to various nebulæ that may still be seen by the telescope in remote parts of the sky; that the gradual falling together of most of the cloudy mass about its center produced the sun, while the planets were formed by the gathering together of much smaller amounts of matter about subordinate centers. The correspondence of rotary motions now observable is regarded as a common inheritance from the slow turning of the

original nebula from which the solar system is supposed to have been evolved; and this theory of the origin of the sun and the planets is consequently called the *nebular hypothesis*. When the scattered parts of the early nebula were gathered together, the larger bodies that they formed are believed to have possessed an excessively high temperature. The sun, being the largest of all, still retains much of its primitive heat. The earth, being smaller, has now cooled to a low temperature on its surface; a large amount of heat is, however, still retained within the earth.

5. Evolution of the atmosphere. In the early youth of the earth, when according to the hypothesis its surface temperatures were high, many substances that might later be condensed at lower temperatures in the liquid ocean or the solid crust, would then exist in the atmosphere. Such an atmosphere would be dense and vaporous; heavy clouds would hang in its upper layers, and drenching rains would fall towards the glowing earth, only to be boiled off again as they approached it; until at last by a long process of slow cooling through untold ages, more and more condensation would take place, reducing the volume of the atmosphere to moderate measures, when only a small share of its original mass would remain. Changes of this kind would take place faster on the smaller planets, slower on the larger ones; and this seems to be the fact in our own system. The moon, a comparatively small body, appears to have lost all its atmosphere. Jupiter, much larger than the earth, appears still to possess a very cloudy atmosphere; and from the great brightness of this planet, astronomers have been led to suppose that its body is still so hot as to be somewhat luminous. The sun, vastly larger than any of the planets, still retains an atmosphere of great volume at excessively high temperatures, which its small neighbors have long ago lost.

Our earth occupies an intermediate position. Some of the more volatile mineral substances in the rock-crust of the earth presumably at an early time made a part of the atmosphere, but all these have long ago left it. Nearly all of the water that must have once been boiled off in the steamy atmosphere of early times has now condensed upon the cooled surface of the earth, forming the deep oceans. Some of the gases themselves, particularly the oxygen of the air, must have been much diminished by combining with the surface rocks of the earth's crust and rusting them.

It is also possible that the early atmosphere has been diminished not only by condensation and combination on the earth, but also by flying away from the earth. If lighter and more active gases, such as hydrogen, ever existed free in the atmosphere, it may be plausibly supposed that they have escaped from the earth's attraction and passed out to open space, to be gradually gathered around larger planets or suns. The absence of even the heavier gases of our atmosphere around smaller bodies, such as the moon, has been thus accounted

for. The atmosphere, at the bottom of which we live, must, therefore, be regarded simply as the thin residual of the much vaster early atmosphere that once surrounded the earth.

6. The future of the atmosphere. We may not only look back into the past; we may peer forward into the future, and speculate as to the further changes still in store for the atmosphere. The earth already having cooled greatly by the comparatively rapid loss of its own heat, the further lowering of temperature on its surface depends chiefly on the slower cooling of the sun. When the sun at last becomes cold and dark, all the water vapor will have forsaken the atmosphere, and our oceans will have frozen solid. The air will be absolutely calm, and all the dust will settle from it, leaving it a pure, clean gas. More of the oxygen will have then combined with the rocks of the earth's crust; possibly nearly all of it may by that time have been withdrawn from the atmosphere; but the nitrogen, the inert element of the air, will remain, little changed from its present amount. We cannot easily imagine any process by which the nitrogen of the atmosphere will be disposed of, unless the surface of the earth becomes so absolutely cold that the gaseous condition should be lost and the nitrogen should condense as a solid on the frozen earth.

The future does not, according to these speculations, appear to have in store so great a change as has occurred in the past. When the earth is cold and the sun dark, the atmosphere will be somewhat thinner than now, but its decrease in volume will not be nearly so great in the future, while the sun cools, as it has been in the past, during the cooling of the earth.

The changes in the condition of the atmosphere, here so briefly reviewed, have required the passage of untold ages of time. All the millions of years during which the earth has already possessed temperatures fitted for the existence of life on its surface, form but a short middle chapter between the much greater duration of its ardent youth, long past, and its cold old age, yet to come. While we may gain some general conception of the changes that have taken place and that are yet in store for the earth, the time measured by these changes passes our comprehension.

7. Composition of the atmosphere. As at present constituted, pure, dry air, from which the dust, water vapor, and carbonic acid have been taken away, consists of oxygen and nitrogen in the proportion of 21 to 79 parts by volume. These two gases are not chemically combined, but are simply mixed together. Their mixture is very perfect, and extraordinarily uniform the world over. Analyses of samples of air collected from all the continents, from many parts of the oceans, from sea-level, from mountain tops, and from lofty balloon voyages show hardly any variation in the proportion of these two chief constituents. This is because the atmosphere is extremely mobile,

and because gases possess the property of spontaneous mixture or diffusion, whereby inequality of composition is soon lost.

The ordinary atmosphere possesses in addition to the oxygen and nitrogen a small part, about three-hundredths of one per cent., of carbonic acid. This varies slightly, being a trifle less by day and in the summer, than by night and in the winter; but the changes of its proportion are extremely minute. There is also a variable quantity of water vapor, sometimes locally amounting to three per cent. of the air by weight, but generally much less. Besides these, there are occasionally minute quantities of accidental constituents, produced by lightning, such as ammonia, nitrous acid, and ozone,[1] in addition to various microscopic solid particles, such as dust from the land, salt from the sea, the pollen and spores of plants, and innumerable organic germs.

Nitrogen, which constitutes the largest part of the atmosphere, is a comparatively rare element in the earth. The probable explanation of its large amount in the atmosphere is found in its chemical inertness. It does not easily combine with other substances, and hence, although a rare element in the earth as a whole, is common in the atmosphere from having been left over at the time when other elements united to form liquid or solid substances.

Oxygen, on the other hand, is one of the commonest substances in the earth. It forms a large proportion of the waters of the ocean and of the superficial rocks of the earth's crust. It constitutes a small share of the atmosphere, not because it was in small quantity in the begining, but presumably because its original abundance was actively reduced by uniting with other substances in chemical compounds. In spite of its having been originally in great quantity, it now makes the smaller part of the atmosphere.

Carbonic acid, a compound of carbon and oxygen, is trifling in amount in the atmosphere and yet is of essential importance to the vital processes of plants, as will be seen in the next sections. It is given off with water vapor and other gases in volcanic eruptions; its carbon is taken in by plants, which in times long past have thus stored up great quantities of carbon in coal beds. Hence the proportion of carbonic acid has probably varied during the evolution of present conditions; though it should not be inferred that all of the carbon now existing as coal was at any one time combined with oxygen, and thus added to the store of carbonic acid in the atmosphere.

In some volcanic districts, carbonic acid is given off plentifully enough to accumulate in the hollows and render the air poisonous. An example of this

[1] Ozone is an allotropic form of oxygen; its molecule consisting of three oxygen atoms, while the ordinary oxygen molecule consists of two atoms. Ozone has a peculiar odor, whence its name. It acts as an oxydizing agent, because it easily gives up one of its atoms, thus returning to the condition of ordinary oxygen. Its presence is generally tested by means of this property; the rate of change of some easily oxidized substance being taken to measure the amount of ozone present at the time. This test, however, is not accurate, as the same change may be caused by other atmospheric impurities.

is found in Death Gulch, a ravine in our Yellowstone National Park, where the proportion of the gas emitted from the ground is sufficient to suffocate animals that stray there.

8. Offices of the atmosphere: relation of oxygen to animals and plants. The peculiar relation of the atmosphere to organic life may be explained by analogy with the case of an ordinary steam engine. A steam engine burns fuel, such as coal or wood, in order to gain energy to do the work that it has to perform. All plants and animals burn fuel, that is, some part of their organic substance, for the same purpose. Engines do their work by the energy of high-pressure steam that has been formed from water and raised to a high temperature by the heat from the fiery combustion of the fuel in a grate or fire-box close to the boiler; and the essential supporter of this fiery combustion is the oxygen of the air. The work performed by animals, such as walking, swimming, flying and everything else in which resistance is overcome, is done by means of the energy gained from a fireless combustion, a slow combination of some of their blood with the oxygen of the atmosphere; the oxygen that fish find dissolved in the water having been gained from the atmosphere above.

All plants also have some work to do; truly a trifle compared to that accomplished by animals, but still properly named work; either in the lifting of the sap from the roots to the leaves, or in the moving of the roots, the tendrils, or the leaves; and like animals, they gain the energy for this work by a slow combustion, a combination of part of their organic substance with the oxygen of the air, which goes on in all their living cells.

In both animals and plants, the combustion here referred to is associated with the process of respiration, corresponding to the draft of the fire in an engine. Respiration includes the inhalation of a certain amount of air, the combination of part of the oxygen in the inhaled air with some of the organic substance of the plant or animal, and the exhalation of the products of combustion with the unused air. The process is much alike in plants and animals, differing rather in quantity than in kind, although performed by very different organs. The products of exhalation are chiefly carbonic acid and water. In the case of animals, these are accompanied by certain noxious organic vapors, and it is from the latter that the unpleasant odor and oppressive feeling arise in poorly ventilated rooms.

9. Relation of carbonic acid to plants. In another way, plants and animals are strongly unlike. This is in respect to the food on which they live. Animals require some organic substance for food, either from plants or from other animals. The food is used to build up their bodies, to repair their waste, or to support the slow combustion that has already been referred to, by

means of which they can do work. The higher plants, on the other hand, live as a rule on inorganic substances, which they derive from two sources. The sap, consisting mostly of water with a small amount of mineral and organic substance dissolved in it, comes from the earth through the roots. It rises through the stem and branches to the leaves, where much of it evaporates. But another part of the food of the higher plants comes from the carbonic acid of the air, and it is in this relation to the atmosphere that plants and animals are so unlike. The carbonic acid of the atmosphere is of no use to animals. It is given out in the exhalation of their breath, just as it is exhaled, truly in very small quantity, from the breathing cells of plants; but in plants the carbonic acid of the atmosphere is taken in by the green cells of the leaves, and under the action of sunshine it is decomposed, the carbon being retained and the oxygen given out. This process, therefore, goes on only in the daytime, and not both day and night, as in the case of breathing. The sap and the carbon gained from the earth and the air, constitute the food of the higher plants. From these, the plant builds up its tissues, repairs its waste, and supplies the fuel for the very gentle combustion that goes on in its breathing cells. The oxygen liberated by the decomposition of the carbonic acid in the green cells goes off to the air, and thus about balances the consumption of oxygen by plants and animals. It therefore appears that the relations of plants and animals to the oxygen and carbonic acid of the atmosphere are in part alike and in part very unlike.

10. The nitrogen of the atmosphere has already been referred to as a very inert element. It does not appear to have any direct use. By increasing the density of the atmosphere, it enables the voice to be heard further than it would be in a thinner gas; it makes flying easier for birds and insects; it makes the wind stronger and more serviceable in turning windmills and blowing the sails of ships; by diluting the oxygen, it diminishes the activity of combustion, which in an atmosphere of pure oxygen would be excessive; but it does not appear to act in any direct way.

Water vapor, the most variable component of the atmosphere, is of extreme importance in many regards. The movement of vapor in the atmosphere constitutes one member in the continuous circulation of the waters of the world, beginning in the evaporation of water from the ocean surface, passing then as vapor, carried by the winds, until, condensing in clouds and falling as rain or snow, it reaches the land or the sea; that part which falls upon the land gathers in streams and rivers running down the slopes of the surface and bearing the waste of the land with it to the sea.

11. Dust. The solid impurities of the atmosphere are of varied nature. Besides organic particles of many kinds, mineral dust is raised into the air by

the winds; the coarser particles soon settle down again, but the finer ones may remain in suspension for months or years. The spray blown by the winds from ocean waves may evaporate, leaving its finely divided salt in the air, when it may be carried many miles before settling or being washed down in rain. Explosive volcanic eruptions also furnish large quantities of dust particles, which may be carried by the outbursting gases high into the upper atmosphere (section 71). The lower air, especially in dusty regions, contains thousands or even hundreds of thousands of particles of one kind or another in a cubic inch. Over the oceans or high in the atmosphere, the air is relatively clean and pure. We shall see reason further on for believing that the dust thus suspended in the atmosphere plays an important part in determining its temperature, as well as in illuminating the sky under sunshine and determining its color; and if certain physical experiments in the laboratory apply also in the greater scale in nature, the dust suspended in the air may be of much use, if not essential, in determining the production of fogs and clouds, and thus of rain.

12. Activities of the atmosphere. The air as a whole may be regarded first as a medium in which a great variety of the minutest forms of organic life are carried about. The germs upon which the decomposition of organic matter depends, and which determine in so many cases the occurrence of disease, are carried easily by the lightest movement of the air. The spores and pollen of plants are widely distributed from their source; the winged seeds of many plants are carried for less distances. Most birds and many insects use the air to support their flight, and occasionally mammals, reptiles, and even fish do the same. Air in motion serves to drive sailing vessels over the sea, to turn the wheels of wind-mills, and to support balloons. The latter use, although at present rare and comparatively dangerous, is probably destined to become of much greater service in the future. As a geological agent, the moving air is of great importance in transporting fine dust from place to place, as well as in carrying the vapor by which rivers are fed, as has already been mentioned. Moreover, the winds blowing over the sea raise waves, which beat upon the shoals and the shores, and grind the land down beneath the level of the waters. The winds drive the waves along, and thus create currents which not only largely determine the distribution of the forms of life in the sea, but also have an extraordinary effect in the distribution of temperature in the air. Finally, it is the oxygen of the air which supports not only the combustion that is so essential in all plants and animals, but also the more active combustion of all engines which are driven by fires, and the combustion which directly or indirectly serves to give light at night. In these many ways, our own life and activities, the life of all plants and animals, the life of the earth itself, all depend upon the atmosphere which lies around us.

CHAPTER II.

EXTENT AND ARRANGEMENT OF THE ATMOSPHERE ABOUT THE EARTH.

13. The geosphere, hydrosphere and atmosphere. The great mass of the earth, solid at least in its outer crust, is for the most part bathed in an ocean of water and is entirely surrounded by an envelope of gases. These three parts of our planet are sometimes named the geosphere, the hydrosphere, and the atmosphere; the latter term being in familiar use, the others being less frequently met. The arrangement of the several parts will be better understood if we recall the physical properties of matter in the three states, solid, liquid and gaseous.

Solid bodies retain a definite form and volume, unless acted upon by some severe strain. The solid crust of the earth is slowly strained and crushed into the uneven form of continents and mountains; the inequalities thus acquired would be retained indefinitely, if it were not for the slow weathering and washing away of their surface; but this process of change is so slow that we need not consider it further in the study of meteorology. For our purposes, the geosphere may be regarded as a rigid spheroid, of somewhat irregular surface.

Liquids tend to retain a definite volume; their free upper surface takes a shape at right angles to the resultant of the forces acting on it, while their form elsewhere depends on that of the body upon which they rest. The liquid hydrosphere is acted on by the centripetal pull of terrestrial gravitation; it settles down in the depressions of the earth's crust, forming the oceans, with a surface standing everywhere at right angles to the force acting on it. If terrestrial gravitation acted alone, the surface of the ocean would take the shape of a sphere; for a sphere alone has a surface everywhere at right angles to a system of forces directed to a single center: but on account of the centrifugal force of the earth's rotation, the ocean's surface is deformed into a slightly flattened or oblate spheroid.[1]

[1] The following simple statement of the problem may serve to explain the deformation of the sphere into the spheroid. Inertia is the resistance that a body opposes to a force that changes the velocity or direction of its motion. There is no especial name given to that manifestation of inertia which comes from a change of velocity; but the inertia developed by a change in the direction of motion is called by the special name, "centrifugal force." It is in no proper sense a force; for a force is that which changes or tends to change the direction or velocity of a body's motion. That form of inertia resistance which is called centrifugal force is manifested in a direction opposite to that of the actual force by which the body's path is changed; and hence in the case of a circular motion, or rotation, in which

The level surface of the sea is naturally taken as the standard of reference in measuring the heights to which the land rises above it, or depths to which the floor of the ocean basins sink below it.

Gases do not retain a definite form or volume. They continually exert an expansive force, tending to increase their volume. If acted on by external forces, they may be compressed into smaller and smaller volume; if free to expand, they will increase in volume and occupy all the space allowed them. The gases of the atmosphere are drawn down on the surface of the sea and land by gravity; the weight of the upper strata compresses the lower ones into greater and greater density; while the upper ones expand to an extreme tenuity. We know nothing by direct observation of the free surface of a gas; and hence cannot well understand how the atmosphere is limited upwards.

14. Dimensions of the earth. The following rough table of dimensions is of interest in this connection:—

Area of earth's surface, ...	197,000,000 square miles.
Volume of earth, ...	256,000,000,000 cubic miles.
Mass of earth, ...	6×10^{21} tons.
Area of ocean, . .	150,000,000 square miles, or $\frac{3}{4}$ of earth's surface.
Volume of ocean, ...	300,000,000 cubic miles, or $\frac{1}{853}$ of earth.
Mass of ocean, ...	18×10^{17} tons, or $\frac{1}{3340}$ of earth.
Mass of atmosphere, ...	5×10^{15} tons, or $\frac{1}{1200000}$ of earth.

15. Pressure of the atmosphere. The level surface of the ocean would everywhere be equally pressed upon by the overlying atmosphere, if there

a body is continually pulled by a centripetal force towards a center, the so-called centrifugal force is manifested outward along the radius.

In the case of any part of the ocean's surface layer, A, Fig. 1, the only force acting on it is terrestrial gravitation, AG. One component of this force, AB, must be expended in overcoming the inertia (centrifugal force), AC, that arises from the continual change in the direction of the body's motion. If AB is expended, only the other component, Ag, can remain; hence it is only the latter component of terrestial gravitation which acts to determine the shape of the ocean's surface. This component is called gravity. It is not directed to the earth's center, but slightly away from the center, towards the pole of the opposite hemisphere from that in which A is situated. Consequently, a plumb line at A will hang in the direction Ag, or vertical; a water surface at A will adjust itself at right angles to Ag, or level. A continuous ocean surface from N to Q will everywhere adjust itself to a system of local gravitative forces, all of which are turned a little away from O towards S. Hence NAQ becomes nAq; or the sphere is changed into the oblate or flattened spheroid.

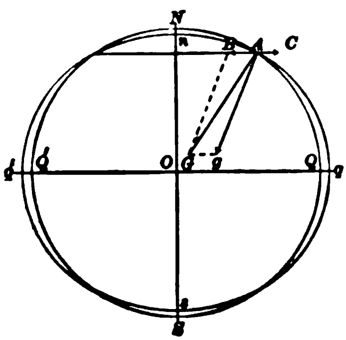


Fig. 1.

were no disturbing forces present by which a greater part of the atmosphere might be accumulated in one region than in another. The surface of the land suffers a less and less pressure, the higher it rises above sea-level. It is the pressure of the atmosphere on the water in a well that raises a column of water in the tube of a pump, where the downward pressure of the air is removed by raising the piston. The height to which water will rise in a pump may, therefore, be used to determine the value of atmospheric pressure. This height is about thirty-four feet, if the experiment is made at sea-level. As the weight of a cubic inch of water is 0.036 of a pound, the pressure of the atmosphere on a square inch of surface at sea-level must be 14.7 pounds, or about a ton on a square foot.

16. Barometers. Water is not heavy enough to be conveniently used in determining atmospheric pressure. The heavier liquid, mercury, is much better adapted to this purpose. If a glass tube, about thirty-two inches long, closed at one end and filled with mercury, be inverted and the open end placed in a dish of mercury, the height at which the mercury will then be held in the tube affords a precise and convenient indication of the pressure of the atmosphere. Mercury being thirteen and a half times heavier than water, or 10,784 times heavier than air, the column of mercury will stand at a height of about thirty inches at sea-level; but the length of the mercury column will be less and less at more and more elevated stations. If the air were uniformly dense at all altitudes, the upper surface of the atmosphere would be found at a height of 10,784 times 30 inches, or about five miles. By attaching a scale to the tube, the height of the mercury may be read, and thus the *barometer* or pressure measure is constructed. It is customary to speak of the pressure of the atmosphere in terms of barometric inches; that is, the height in inches of the column of mercury that the pressure of the atmosphere sustains in the tube at any time. The precise measure of what is called one atmosphere of pressure in these units is 29.905; the mercury having a temperature of 32°, and the observations being reduced to the latitude of London (see Sect. 101): this equals a pressure of 760.00 mm. at latitude 45°. Further account of the construction of the barometer and its use will be found in Chapter VI.

A thousand cubic feet of dry air—that is, the contents of a cubic room measuring ten feet on a side—at a temperature of freezing (32° Fahrenheit) and under a pressure of 30 barometric inches, or 30 inches of mercury as it is commonly called, weighs 75.29 pounds. If its volume be kept constant, its expansive force will increase by $\frac{1}{490}$ of the expansive force at freezing for every rise of one degree Fahr. in temperature; but we have little to do with this condition in meteorology. If, as is much more commonly the case, the pressure upon the air remains constant, it will expand as its temperature rises;

its volume increasing by $\frac{1}{491}$ or 0.002 of the volume at freezing for every rise of one degree Fahr. (or by $\frac{1}{273}$ for a rise of one degree Centigrade).

17. Downward pressure of the ocean. Return now to the case of the level ocean, on which the atmosphere everywhere exerts a uniform pressure. If we descend about 33 feet into the salt waters of the ocean, we may imagine a surface there, concentric with the outer spheroidal surface of the ocean, on which the pressure will everywhere be two atmospheres. At a depth of 66 feet, the pressure will be three atmospheres; and so on to the bottom. Such imaginary surfaces are called *isobaric*, from having equal pressure on all parts. At the average depth of the ocean, or two miles, the pressure will be 320 atmospheres; at the greater depths of the deepest parts of the oceans, between four and five miles, the pressure rises to seven or eight hundred atmospheres.

Water, however, is so nearly incompressible that even under these enormous pressures, its density is not greatly increased. The water at the bottom of the great oceans is only about as much denser than the surface water, as the latter is denser than fresh water. Any substance that is heavy enough to sink rapidly below the surface of the sea will sink all the way to the bottom.

18. Isobaric surfaces in the atmosphere. The case of the atmosphere is very different. It is highly elastic, and hence must be much denser at the bottom than near the top. If we ascend about 900 feet from sea-level, the barometric pressure there will be reduced from 30 to 29 inches. At this height we may imagine a level spheroidal isobaric surface, concentric with that of the ocean, on which the pressure of the overlying atmosphere is everywhere 29 inches. How high must one ascend to reach a second isobaric surface on which the pressure would be 28 inches? If the height of the surface of 29 inches is 900 feet, the height of the second surface above the first must be $\frac{29}{28}$ of 900; or 932 feet: for the volumes of gases are known to increase as the pressure by which they are confined decreases. The height of the third surface, of 27 inches pressure, would be $\frac{29}{27}$ of 900, or 967 feet above the second surface; and in this way we may continue to calculate the altitude at which any pressure would be found. The pressure of 26 inches would thus be found at a height of 3,800 feet above sea-level. If the rule were followed to the last case, it would lead us to say that the height at which the surface of no pressure — that is, the upper surface of the atmosphere — would be found, would be $\frac{29}{0}$ times 900 feet above the surface of one inch pressure, and this would be at an infinite distance above it; but it is not probable that the rule by which volumes and pressure are correlated can be fairly applied to such extreme cases. Moreover, the known decrease of temperature in the upper air would somewhat reduce the measures here given. It must suffice to say that

the air becomes thinner and thinner as we ascend above sea-level; that the successive isobaric surfaces are separated by greater and greater distances; but of the absolute termination of the atmosphere we can say nothing.

19. Vertical decrease of pressure in the atmosphere. It is possible, however, to calculate with a close approach to accuracy the height at which any given pressure will be found; or, conversely, the pressure corresponding to any given height. Allowance must be made in this calculation for the decrease of temperature at the rate of 1° in 300 feet in ascending above sea-level; and for certain other minor corrections; when this is done, we find the results given in the following table:—

Pressure.	Altitude.
30 inches.	0 feet.
29	910
28	1,850
27	2,820
26	3,820
25	4,850
24	5,910
23	7,010
22	8,150
21	9,330
20	10,550
18	13,170
16	16,000

20. Height of the atmosphere. In the older books on meteorology, the height stated for the atmosphere was 45 or 50 miles; this being the altitude at which no significant barometric pressure would be encountered; at a height of 30 miles the pressure is only half a hundredth of an inch. Observations on the duration of twilight — that is, of the perceptible sunlight which is turned from its direct course by the action of the atmosphere after the sun has set — gave about the same measures; but this depends manifestly upon the delicacy of the observations by which the duration of twilight is determined; if our eyes were sharper, twilight might be perceived longer. Hence in this case, as in the other, all that can be said is that at a height of about 50 miles, the air becomes excessively thin, incapable of producing significant pressures, or of deflecting perceptible amounts of sunlight.

Observations of meteors, however, give much greater dimensions for the height of the atmosphere. Meteors are small solid bodies, flying rapidly through space, and sometimes entering the earth's atmosphere. Their velocity is so great that they generate heat enough by the compression of the air in

their path to render them luminous, and even to disintegrate them before they reach the bottom of the atmosphere. Only the larger ones reach the ground unconsumed. Meteors have sometimes been seen from several places by different observers, and their apparent paths among the stars determined closely enough to enable one afterwards to calculate the angular altitude of each observer's line of sight above the horizon. The height at which the lines of sight intersect may then be easily determined; and in this way it is learned that meteors become visible at heights even greater than 100 miles. Although it is difficult to conceive of the excessive tenuity of the air at such heights, we are constrained to believe that it exists there; how much further the faintest traces of the atmosphere may extend must be left to future discovery.

Valuable observations of meteors may be made by any persons who can record the time of appearance accurately and whose knowledge of the constellations is sufficient to identify the stars easily; or, even without this knowledge, by persons who will take care to notice the track of a meteor past some fixed objects, whose direction and angular altitude may afterwards be measured by surveyor's instruments. Such observations, reported to the central office of any of our state weather services, may be compared with good results. A meteor seen over New England on September 6, 1886, and reported by several observers of the New England Meteorological Society, was determined by Professor Newton of Yale College to have become visible at an altitude of 90 miles over northwestern Vermont, and to have disappeared at an altitude of 25 miles over southeastern New Hampshire.

We know little of the vast upward expanse of the atmosphere. The rays from the sun and stars enter through it; hence it must be excessively thin and pure. Meteors dart into it, and, if heavier than a few ounces weight, in most cases fall to the earth. But the highest mountains do not rise six miles above sea-level, and their upper slopes are deserts of rock and snow. The highest flight of a balloon, nearly seven miles, reached temperatures so low and air so thin that the balloonists fainted. All the clouds of those lofty regions consist of minute ice crystals; but they are seldom, if ever, measured at heights greater than eight or nine miles. It is only the lower part of the atmosphere that is of sufficient density and of a high enough temperature for the easy maintenance of life, as it has been developed on the earth. The greater density of the lower strata has already been explained as a result from the pressure of the upper strata; we may next inquire as to the control of the temperature of the atmosphere.

CHAPTER III.

THE CONTROL OF ATMOSPHERIC TEMPERATURES BY THE SUN.

21. Sources of heat. The only sources of heat on which the temperature of the atmosphere may be conceived to depend are the sun, the stars and the earth.

The earth is still very hot within; but its hot interior mass is so well sealed over by a non-conducting crust, that very little heat escapes outward through it. Volcanic eruptions occasionally carry some of the glowing rocks from the interior up to the surface in a spasmodic manner; but these eruptions are manifestly too exceptional for further mention. Moreover, if the temperature of the air over the earth's surface depended on conduction from the earth's interior, we should not know how to explain the changes of temperature from the equator to the poles, or from summer to winter.

The stars are as hot as the sun, and they are innumerable; but their distance is so great that they control our temperature as little as they control our light. Moreover, they shine from all parts of the sky; hence, if our temperature were said to depend on star beams, we should not know how to explain the changes of temperature from day to night.

The sun is manifestly the ruler of temperatures on the earth's surface. The changes of temperature from equator to pole, from summer to winter, from day to night, all follow the changes in the intensity of sunshine. There can be no doubt in an explanation where variations in the effects follow so precisely the variations in their presumed cause.

When we come to explain how it is that the sun controls terrestrial temperature, it is necessary to go slowly, if we would gain a clear understanding of the process.

22. Nature of heat. It must be now recalled from the study of physics, that heat is not a thing in itself, but simply the energy of the molecular motion of any material substance. If two masses of lead be struck violently together, they become hotter than before. Iron and steel are not so well fitted for this experiment, for, by reason of their elasticity, their energy of impact is mostly expended in producing a rebound. Lead being relatively inelastic, it is supposed that the energy of the impact is expended in exciting the molecules of which the masses are composed to a more active motion; and we recognize this more active motion in the higher temperature of the bodies. It must be clearly understood that, in speaking of molecules, or of heat as the

energy of molecular motion, we are speaking of things that have never been seen; of phenomena that have never been observed. It is, nevertheless, reasonable to believe in molecules and in the mechanical theory of heat, as it is called, because by the acceptance of such hypotheses, we are enabled to explain and correlate a great variety of well ascertained facts, that have otherwise found no explanation.

Heat, then, is believed for good reasons to be the energy of molecular motion. When the surface of the ground and the lower layers of air become warmer under the rays of the morning sun, we believe that their molecules have been excited to greater velocity of movement: but then the question arises—how can the velocity of their molecules be affected by the sun, which is distant from the earth by ninety-two million miles! The space between the planets and the sun must be empty; otherwise, the planets could not maintain a constant distance from the sun; they would approach it on an inward spiral path, moving faster as they neared it, and thus accomplishing an annual revolution in a decreasing number of days. This is not the case, as far as observations go. We must, therefore, suppose space to be essentially free of resisting matter; if any medium is there for the propagation of energy from one mass to another, it must be of properties quite unlike those of the gross forms of matter that we know on the earth, rarer even than the extremely tenuous upper strata of the atmosphere. Yet some continuous medium must be supposed to exist all through space; for without it we cannot advance towards an understanding of the process by which the light of the stars reaches us; or by which the sun controls our temperature; or by which a hot ball of iron can become cool even though suspended in a vacuum.

23. Explanation by hypothesis. Here, as in the mechanical theory of heat, we must have recourse to some hypothesis; and our faith in the hypothesis should be measured simply by its success in explaining facts of observation. It is commonly the case in the progress of science that various hypotheses are successively advanced in the attempt to explain facts of observation; the hypothesis which best accounts for all the facts will come, in time, to be generally accepted. For example, many facts are known concerning the motions of bodies; the falling of any object towards the earth; the movements of the moons around the planets, and of the planets around the sun; and even the revolution of the two components of a double star around a common center. All these facts of motion may be explained by the hypothesis that every mass of matter attracts every other mass with a force directly proportional to the product of their masses, and inversely proportional to the square of the distance between the two. This hypothesis is so successful in explaining all relevant facts, that it has come to be universally accepted. The force of attraction is called gravitation; the statement of the manner in which it

varies is called the law of gravitation. The discovery and establishment of this law is the chief glory of the immortal Newton.

The hypothesis by which we seek to explain the loss of heat from a hot ball suspended in a vacuum, or from the sun standing in empty space, must now be briefly outlined. It may be named the hypothesis of radiant energy.

24. Radiant energy. It has been supposed that in spite of the apparent emptiness of space, as far as molecular matter is concerned, it is nevertheless filled with an all-pervading medium of perfect continuity, excessive rarity and extraordinary elasticity. Any disturbance, such as the agitation of the molecules of ordinary matter, at once imparts a disturbance to the imagined medium; and then, much in the same way as waves spread out from a point where a stone falls into a body of water, so waves of disturbance spread out or radiate through the imagined medium in all directions from the center of excitement.

Whether the hypothetical medium be named the luminiferous ether, or whether it takes a name from its property of propagating electro-magnetic disturbances, this matters nothing to us now; all that the hypothesis of radiant energy demands is that molecular disturbance should spread away or radiate by a wave-like motion or undulation in some space-pervading medium. Hence the name, *undulatory hypothesis*, originally employed, when the idea of the undulatory nature of light was first introduced by Huyghens in the seventeenth century, in distinction to the corpuscular hypothesis of Newton, which assumed that the light from the sun consisted of a shower of minute corpuscles.

The undulations of radiant energy travel away in straight lines in all directions from their source at the enormous velocity of nearly two hundred thousand miles a second. They require only about eight minutes to span the space from the sun to the earth. The dimensions of the undulations are excessively minute; those given out from the sun varying in length between 0.00270 and 0.00029 millimeter, or 0.00011 and 0.00001 inch. Their undulation is incredibly rapid; reaching several hundred million million undulations in a second, the finer waves swinging the more quickly. They may be likened to the much larger waves of sound excited in the atmosphere by an orchestra. The trombones and bassoons produce long waves of relatively slow undulation; the fifes and the high notes of the violins excite finer waves of relatively rapid undulation; the louder notes excite waves of greater breadth of swing, or amplitude; but they all travel away at a uniform velocity. On encountering any object, they spend their energy upon it; that is, they set it into vibration; thus the waves from one tuning-fork may set up vibrations in another. If the waves of sound disturb the delicate mechanism of the ear, they give us the

sensation of hearing. In a very similar way, the waves of radiant energy, varying in period and amplitude of undulation, move on from any exciting source in straight lines or rays at a uniform velocity as long as they pass through what we call empty space; but if the waves encounter sensible matter, some of the energy of undulations is imparted to its molecules, and the velocity of molecular movement is increased; that is, the mass that the molecules constitute is heated. A careful distinction should be drawn between radiant energy, whose nature is essentially undulatory, and molecular energy or heat, which is characterized by a confused molecular agitation.

25. Radiation from the sun: insolation. The sun is an enormous globe of excessively hot matter; many times hotter at its cloudy surface than the hottest furnace, and presumably hotter still within. Its mass is so great that in spite of its active radiation of energy, by which its heat is reduced, it will be hot for ages to come. Its diameter is about 880,000 miles; if the earth were at the center of the sun, and the moon were revolving about the earth at its present distance of 240,000 miles, there would still be a solar shell outside of the moon 200,000 miles thick.

The radiant energy or radiation emitted by the sun is conveniently given the special name of *insolation*. It varies greatly in wave-length and period of undulation. It flies away in all directions at an incredible velocity. The greater part of it goes on and on for ages through the void of the universe, constantly becoming fainter as it embraces wider and wider spheres of action; only a minute part of the emitted insolation encounters a planet on its way through space.

The heat emitted from a small area of the sun's surface has been compared with that given out from an equal area of melted steel in a Bessemer furnace; the ratio being 87 to 1 in favor of the sun. The heat received from the sun's rays falling vertically and unobstructed on a square mile of the earth's surface would warm 750 tons of water from the freezing to the boiling point in a minute. The whole amount of heat received from the sun on the earth in a minute would warm 37,000,000,000 tons of water by the same amount. The heat thus received would suffice to melt a layer of ice about 160 feet thick over the whole earth in a year; this is several thousandfold greater than that received from the earth's interior; and yet the earth receives only one two-billionth part of the heat given out by the sun! The other planets receive similar minute fractions; the rest is "wasted"; that is, we do not see that it is applied to any particular purpose. Upon the trifling share of insolation received by the earth depend all our activities, except those of telluric origin, such as earthquakes and volcanoes; and those of lunar origin, such as the oceanic tides. The winds, the ocean currents and all the processes of life depend on energy received from the sun.

Whatever one may feel about the correctness of the undulatory theory of radiant energy and however difficult it may be to grasp its fundamental conditions, it must be recognized as more nearly explaining the facts with which it is concerned than any other theory that has been proposed. Its essential feature of undulation is universally accepted by physicists, although the nature of the medium by which the undulations are transmitted is by no means understood.

26. Astronomical relations of sun and earth. Having now considered the method by which the sun's heat is transformed into radiant energy and thus propagated outward in all directions, so that the earth is constantly in receipt of a small fraction of the total, we must next examine the position of the earth with respect to the sun at different times of the year, so as to understand clearly how the incident insolation is distributed over its surface; for on this distribution depends the division of the earth into zones and all the changes of the seasons.

The earth moves around the sun in a nearly circular orbit. The actual form of the orbit is an ellipse, but of so faint eccentricity that, the sun being in one focus, the earth is only three million miles nearer the sun at one time than another. The time of nearest approach, called perihelion, occurs conveniently for our memory on New Year's Day; and the time of greatest distance, called aphelion, on July 1. It may be seen at once from this that it cannot be on our distance from the sun that the winter and summer of the northern hemisphere depend.

27. Distribution of insolation over the earth. The axis of the earth, on which it turns once a day, does not stand vertical to the plane of its orbit; but is inclined twenty-three and a half degrees from the vertical, and in such a direction as to turn the north pole away from the sun on the 21st of December; that is, ten days before the time of perihelion. As the earth moves around the orbit, the axis always stands parallel to itself; hence on June 20, the north pole will be turned toward the sun. These dates are called the *solstices*, because the sun then stands farthest south or north of the plane of the earth's equator. It follows from this that the amount of insolation received at different latitudes will vary greatly during the year; first, because the inclination of the sun's rays to the horizon varies; second, because the diurnal duration of sunshine, or the part of the twenty-four hours in which the sun stands above the horizon, varies. The change of seasons is thus determined.

The noon altitude of the sun and the length of the day at any latitude are best illustrated by fitting a paper ring around a globe in the attitude of a great circle; the globe, with the axis properly tilted, being carried around a

curve to represent the orbit, and the paper ring being always adjusted at right angles to a line from an imaginary sun within the orbit. The ring will then separate the light and dark, or day and night halves of the earth, and may therefore be called the twilight circle. If a line is drawn through the sun at right angles to the solsticial line, it will intersect the orbit at two points; and while the earth occupies either of these points, the twilight circle passes through the poles, and the days and nights are everywhere equal. The points are, therefore, called the equinoxes, and are passed on March 20 and September 22. It should be noted that the line defining the equinoxes does not cut the orbit in halves, and that seven days longer time is spent in passing from the vernal equinox through aphelion to the autumnal equinox than from the autumnal through perihelion to the vernal.

At other times than the equinoxes, day and night are unequal, because the twilight circle then cuts the latitude circles unsymmetrically. This effect is strongest at the solstices, when the twilight circle is most oblique to the latitude circles; but the equator always has equal days and nights, because, being a great circle, it must always be bisected by the twilight circle. A little practice with a globe should make this plain.

The altitude of the sun over an observer's horizon is always greatest at noon. The noon altitude will be 90° at some point within the tropics throughout the year; elsewhere even the noon rays fall obliquely, and their effect weakens as they spread over a greater surface than their cross section. This may be likened to the noon sunshine in winter on a north and south road passing over a hill; the snow on the southern slope, receiving the insolation more nearly at right angles to the surface, may be melted, while it remains frozen on the northern slope, where the insolation falls more obliquely.

The proportionate amounts of insolation received in a single day at different latitudes and at different times of the year have been carefully calculated, and are presented in abstract in the following table. The unit of these measures is the amount of insolation received at the equator on the day of the vernal equinox, March 20, when the sun passes through the plane of the earth's equator on its way into the northern hemisphere of the sky.

Latitude	0°	+ 20°	+ 40°	+ 60°	+ 90°	− 90°
March 20	1.000	0.934	0.763	0.499	0.000	0.000
June 21	0.881	1 040	1.103	1.090	1.202	0.000
September 22	0.984	0 938	0.760	0.499	0.000	0.000
December 21	0.942	0 679	0.352	0.000	0 000	1.284
Annual Total	347	329	274	197	143	143

The same data are also presented in graphic form in Fig. 2; the latitude being given on the left margin, the time of year on the right margin, and the

value of insolation being indicated by a vertical measure from the plane of the two margins.

The most important lessons of the table and diagram are as follows:— (1) On the equinoxes, the greatest amount of insolation is received at the equator, where the day is twelve hours long and the sun passes through the zenith at noon; on this day at higher latitudes in either hemisphere, although the day is still twelve hours long, the sun does not reach the zenith and the value of insolation progressively diminishes; at the poles, where the rays pass tangent to the surface of the earth (except for a slight bending by atmospheric refraction), no insolation is received. (2) On our summer solstice,

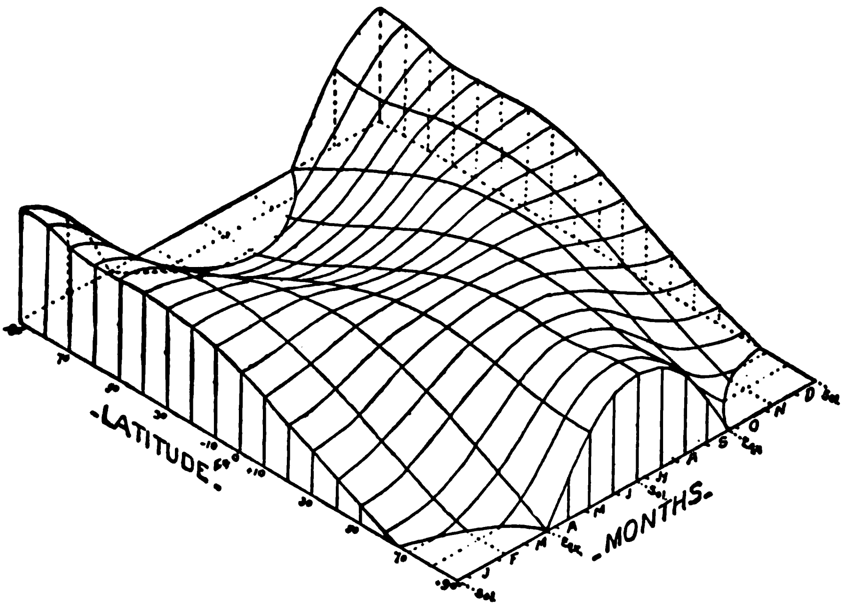


FIG. 2.

June 21, the equator still has a day twelve hours long, but the sun does not reach the zenith then, and hence less insolation is received there on this date than on the equinoxes. Southern latitudes have still more oblique sunshine and a shorter and shorter day, until beyond latitude 66½° the night is twenty-four hours long, and the insolation for the whole southern frigid zone is zero. North of the equator, the sun reaches the zenith over latitude 23½°, and as the day is there more than twelve hours long, the amount of insolation received at this latitude on the solstice is greater than that received at the equator on the equinox. Going further north, there is for a time an increase in the value of the diurnal insolation, because the loss from the lower noon altitude of the

sun is overcome by the gain from the greater length of the day; a maximum value being reached about latitude 40°; then there is a slow weakening, as the decreasing noon altitude more than compensates for the continued gain in the length of the day, the minimum being found just beyond latitude 60°; from this latitude, northward, the increasing length of the day and the presence of the sun above the horizon during the whole twenty-four hours cause an increase in the value of diurnal insolation, reaching a strong maximum at the pole, where the amount received is decidedly greater than at the equator on any day of the year. (3) On our winter solstice, December 21, the relations of the northern and southern hemispheres are inverted; and the south pole then receives even a greater measure of insolation than the north pole received six months before, because the earth is then near perihelion. The limits of the zones are thus seen to be related to the distribution of insolation.

It should be noted that the distribution of insolation depends largely on the length of the day as well as on the altitude reached by the sun. It also appears that the known distribution of temperature on the earth does not closely follow the distribution of insolation; for, if so, the north pole should have a relatively high temperature on June 21; and the south pole should have even a higher temperature on December 21. The explanation of this discrepancy will be found in Section 91.

28. Action of insolation on the earth. An important step may now be made in considering the action of insolation on the earth. We have learned the nature of this form of energy; we have seen its distribution over the earth at different seasons; the effects that it produces come next in order.

It must be carefully borne in mind that radiant energy, while on its way from the sun to the earth, is not heat. It was excited by the heat of the sun, where it was emitted; and it will produce heat when its energy is acquired or absorbed by the substances of the earth; but until thus absorbed, it must be regarded only as a special form of energy; a shower of almost immeasurably rapid undulations, varying in period and amplitude, but constant in velocity of propagation.

29. Reflection. When radiant energy from any source encounters sensible matter, it may be turned back or reflected; it may be passed on or transmitted; or it may be expended in adding to the molecular energy of the body; that is, the energy is absorbed and the body is heated. Burnished silver is the best reflector known; it turns back nearly 98 per cent. of all rays incident upon it; and this in a most systematic manner, so that the angle of reflection equals the angle of incidence. The reflected rays depart without loss of energy, and the reflecting body is not warmed by them. A good reflector can neither be heated easily by absorption, nor be cooled easily by its own radiation.

Snow and water, whether on the surface of the earth or in the clouds, are among the best natural reflectors; much of the insolation that falls on them is turned back into space, and the earth gains nothing from it.

30. Transmission. Rock salt is the best transmitter of all solid substances. It is much better in this respect than glass, which absorbs many of the finer and coarser waves. Such a substance is said to be *diathermanous*, or to possess the quality of *diathermance*; these terms, literally meaning "open to heat," having been introduced when radiant energy and heat were confounded. The temperature of a transmitter is unchanged by the radiant energy that passes through it. It can be warmed only by the energy of the few waves that it absorbs; hence like a reflector it must warm slowly, even in the full glare of sunshine.

The gases of the atmosphere are almost perfect transmitters. Any given thin layer of air retains very little of the insolation incident upon it; it reflects none; nearly all passes through it. It can, therefore, warm very slowly. The pure water of the ocean is also a comparatively good transmitter; but water is much denser than air, and the sun's rays are so weakened by the slight absorption of successive layers of water, every one of which takes a little of the passing energy, that at greater depths than a few hundred fathoms, sunshine must be practically imperceptible. It is curious to discover in this connection that many animals inhabiting the deeper parts of the ocean have well developed eyes, and are brightly colored; and hence we must suppose that light from some source reaches them. Animals dwelling in dark caverns do not develop their eyes, and are white or gray; hence caverns must be darker than the bottom of the ocean. The phosphorescence of many marine animals may serve to illuminate the ocean bottom.

31. Absorption. Carbon is an almost perfect absorber. It reflects only a trifle of incident insolation and absorbs the rest; it is, therefore, rapidly heated. The surface of the greater part of the land, being a poor reflector of radiant energy and transmitting none beneath the surface, absorbs nearly all that falls on it, and warms rapidly under sunshine.

32. Various effects of absorption. A brief digression must be made here to refer to certain other effects produced by the absorption of insolation. If the solar waves fall on certain substances, such as those used on photographic plates, chemical changes of composition follow; but these changes are not to be regarded as the work of the absorbed rays. Substances thus affected may be regarded as so many springs, bent by previous chemical reactions into constrained positions, from which they would gladly free themselves if they could but make a beginning. The excitement caused among the molecules by the absorption of sunshine or of radiant energy from any other sufficient

source merely releases the springs; and the consequent rearrangement of composition is the work of chemical attractions then allowed to operate. Certain wave-lengths are more effective than others in touching off certain substances. In ordinary photographic processes, where salts of silver are employed, the coarser solar rays have little effect; the finer rays are most useful; and these are for this reason sometimes called actinic rays. Sometimes other substances are employed, on which the coarser waves are more active.

Most animals have developed special organs, which we call eyes, that are particularly sensitive to the action of radiant energy of certain wave-lengths. In the human eye, no effect is produced by waves of greater length than 0.00075 mm. or of less length than 0.00036 mm.; but rays of intermediate wave-lengths, if of sufficient strength or amplitude, produce a sensation which we call light. If the coarser waves preponderate, we call the light red; if the finer ones are the stronger, we call the light blue; the intermediate ones corresponding to the various colors of the spectrum. Rays perceptible by the eye are therefore called rays of light, or optical rays.

As has already been stated, when rays of any wave-length are absorbed by ordinary substances, heat is produced. This is true of rays of much coarser wave-length than can be perceived by the eye; hence, it has been common to speak of such as "dark heat rays."

It has happened that the progress of science has been in an order almost the opposite of that in which the subject of radiant energy is here presented. The radiation of light was first studied; then "radiant heat" was investigated. It was the custom for a time to name the different kinds of solar undulations after the effects that they produced; those of intermediate length were called "light rays"; the finer ones, "chemical or actinic rays"; the coarser ones, "heat rays." With this nomenclature, it was implied that the differences between light, heat and chemical action were inherent in the rays themselves. As now understood, the differences depend on the nature of the substance by which the rays are absorbed. For this reason, no proper appreciation of the dependence of the earth's temperature on solar energy can be gained until it is clearly understood that, while the rays are on the way across the emptiness of space, their waves differ only in length, in period of undulation and in amplitude. None of the rays should in any proper sense be then called light rays, heat rays or chemical rays. They should be named only according to their inherent peculiarities, namely, their length or their period of vibration; and not according to their variable effects.

If a beam of solar rays be split up by refraction in passing through a prism, and thus sorted out into a spectrum according to the wave-lengths, the rays of different wave-length can be examined in turn. Taking any special ray, for example, one whose length is 0.00050 mm., it will, if received in the eye, cause the sensation of green light; if a thermometer bulb is placed before

it, a slight rise of temperature will show that the energy of the ray has been converted into heat; certain specially prepared substances would be excited by the same ray to rearrange themselves chemically. How impossible is it then to base a terminology suitable for the rays of insolation on the effects that they produce upon various substances.

While the sensory nerves of the body can detect no differences among rays of different wave-length, except in so far as they produce different amounts of heat, the nerves of the eye have, by means of their "rods and cones," become able to distinguish between the action of rays of different wave-lengths, and thus to discriminate what we call *colors*.

33. Actinometry. It follows as a corollary from what has just been stated that if we wish to gain a measure of the energy brought to us by the rays of the sun, this can best be done by means of an instrument in which some good absorber, such as carbon, is allowed to absorb all the incident insolation; then if the section of the absorbed beam is measured, if the weight of the absorber is known, and the rise of temperature that it suffers in a given time is determined, a measure of the value of insolation can be gained and compared with other measures. Instruments of this kind are called *actinometers*. They have been used by various observers, but by none more successfully than by Professor S. P. Langley, now Secretary of the Smithsonian Institution. By means of observations made when the sun is high and low in the sky, or from low-level stations and from mountain tops, allowance has been fairly made for the loss of insolation in the atmosphere; and it is thus found that a solar ray of one square centimeter in cross section will raise the temperature of a gram of water nearly three degrees centigrade in one minute. The amount of heat or other form of energy needed to raise the temperature of a gram of water one centigrade degree is called a small *calorie*.[1]

A solar ray of one square centimeter in cross section will therefore yield if totally absorbed, nearly three small calories in a minute. This is called the "solar constant." The most energetic part of the solar spectrum is included within the so-called "light" rays, the infra-red and the ultra-violet rays being of relatively small intensity; and there is good reason for thinking that our eyes have learned to perceive the stronger rays by very reason of their strength having made them the most easily perceptible.

34. Radiation from the earth. All bodies emit radiant energy in a greater or less degree. The sun is the most energetic radiator that we know of, unless an exception be made in favor of some of the stars; but with those we are not here concerned.

[1] Sometimes the unit of weight is taken as a kilogram; the amount of heat in that case is called the large calorie. The English unit of heat is generally based on the pound and the Fahrenheit degree.

Not only the sun, but also the air, the ocean and the land emit radiations. As these bodies can only be seen when illuminated by some source of light, we may infer that the wave-length of their own rays is greater than that of the visible rays; and this has been shown to be the case by Langley, who has measured their dimensions by direct experiment, and found them to be much coarser than any discovered in the sun's rays.

It is by the emission of these extremely coarse rays that a balance is maintained with the absorbed insolation, and thus the temperature of the earth's surface is held at an average, intermediate value. In day-time and especially on clear summer days, the temperature may rise to a comparatively high degree, when insolation is at its maximum value; at night and especially in clear winter nights, the temperature may fall to a very low minimum under the almost unimpeded action of terrestrial radiation. But these high and low temperatures are simply oscillations about a mean value, dependent on the balanced action of absorption and emission of radiant energy by the earth.

Recalling that the air, the water and the land act differently on the insolation incident upon them, it may now be added that their activity in the emission of radiant energy corresponds closely to their activity in absorption. This is the case with all bodies. Indeed, were it not so, those whose power of absorption exceeded their radiating power would become continuously hotter and hotter. With these various preliminary facts in mind, we may next undertake the explanation of the changes of temperature experienced by the air, the water and the land during the changes from day to night, or from summer to winter.

35. Absorption and radiation by the atmosphere. The upper air does not vary much from its mean temperature, as far as the few observations made at great altitudes give us information. This is because it is a poor absorber, and hence also a poor radiator. Insolation passes through it almost unimpeded during the day; and at night its temperature is slowly reduced, because so little of its heat is expended in exciting radiation.

The case is somewhat different with the lower air, particularly over the land; for there the numerous dust particles aid both diurnal absorption and nocturnal radiation, and the land-surface helps to warm the air by day and to cool it by night. The dust particles act as absorbers during the day, and thus become centers of heat for the surrounding air; while at night they are effective radiators, and then reduce the temperature of the air about them. The land-surface by day reflects a considerable share of insolation, and thus causes it to pass a second time through the air; it also emits radiation more actively as it rises in temperature, and thus aids in warming the air. At night the land cools to a low temperature, and the air near it is cooled by radiation to its cold surface. If the air becomes very dusty, as in desert regions, or

activity in radiation corresponds to activity in absorption

smoky, as in the neighborhood of forest fires, or cloudy, as in stormy weather, it may be that the lower strata are shielded from warming by day and from cooling by night by the many particles above them. Air of ordinary cleanness may, therefore, be expected to possess a greater and greater range of temperature as we descend towards the earth, an example of diurnal temperatures for the lower air in clear weather being given in Fig. 10 *a*; but very dusty, smoky or cloudy air must have a level of maximum range of temperature at some height over the earth's surface, and a less range both above and below this height. An illustration of the relatively constant temperature in the lower air during a spell of cloudy weather is given in Fig. 10 *b*.

A practical application of this principle is seen in the method commonly adopted in protecting tender plants from freezing in early or late frosts by building smoky fires about them, and thus providing them with a cover of smoke during the night (Sect. 187). In such cases no frost will be formed below the smoke, if it is dense enough, while the frost may be severe on the surrounding unprotected ground. It may be expected that if a series of observations of temperature were taken at different heights in a dense layer of smoke, a greater nocturnal fall of temperature would be found near the top than at the bottom.

36. Vertical temperature gradient. The general distribution and variation of temperature in the atmosphere may be simply illustrated in a diagram (Fig. 3), in which the vertical scale, *OY*, represents altitude, being divided into thousands of feet (the altitude where pressures of 29, 28, 27, etc., inches occur, is marked according to the table of Sect. 19); and the horizontal scale, *OX*, represents temperature, on the Fahrenheit scale.[1] Many observations on mountains and in balloons have determined that on the average the temperature of the atmosphere diminishes 1° Fahr. for every three hundred feet of elevation. This relation is indicated for a station at sea-level whose mean temperature is 65°, by the oblique line, *AB*, whose inclination is such that for every rise of three hundred feet on the scale of

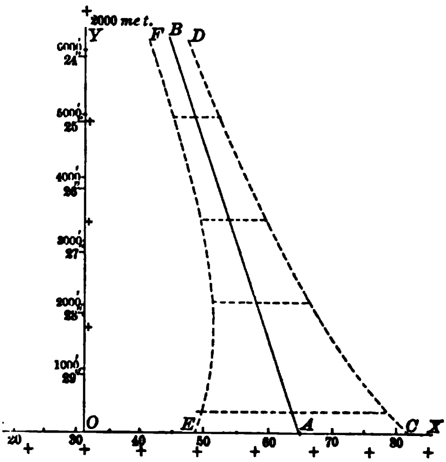


Fig. 3.

[1] The scale of the several diagrams of this kind is constant through the book. Small crosses to the right of the vertical scale mark spaces of 500 meters; similar crosses beneath the horizontal scale indicate even 5° on the centigrade thermometer.

altitude, it shows a decrease of one degree on the scale of temperature. The rate of vertical decrease of temperature, thus expressed either as a numerical ratio or by graphic method in a line, is called the vertical temperature gradient.

From the explanation of the preceding section, we may now add to the diagram a series of horizontal lines, representing the diurnal range of temperature in the air at various altitudes, the middle points of the lines lying on the line of the average vertical temperature gradient. In the upper air the range lines are short; in the lower air, especially over the land, they are longer. Dotted lines, *CD*, *EF*, connecting the ends of the horizontal lines, serve to indicate the vertical decrease of temperature in the atmosphere at times of highest and lowest diurnal temperatures; and as *CD* and *EF* are differently curved, it follows that the vertical temperature gradient is a variable quantity.[1] Its values at different times will be later shown to exert a marked control on atmospheric processes.

The direct rays of insolation arriving at sea-level are weakened by absorption and other losses on their way through the atmosphere. In cloudy weather the largest part of the insolation is detained in the atmosphere; under the densest fogs of London, hardly any perceptible rays from the sun reach the ground. In clear weather a vertical ray passing through the least thickness of air reaches sea-level with about three-quarters of the intensity that it is estimated to possess outside of the atmosphere; but the loss thus suffered is partly made up by the arrival of indirect rays that come from the open sky, having been turned from their initial paths by dust or cloud particles. When the sun is below the zenith, the loss of intensity, as the rays pass obliquely through the atmosphere, is much greater than that suffered by a vertical ray; and at sunset the sun may even be observed by the unprotected eye. While any small volume of air detains only a minute part of the incident insolation, the entire thickness of the atmosphere withholds a considerable share of insolation from the earth. We have now to examine the effects produced by the rays that penetrate to the sea and land.

37. Absorption and radiation by the ocean. The surface layer of the ocean, with which we are concerned in meteorology, is like the air in being slow to change its temperature, but for different reasons. This is a matter of so great importance in determining the climate of many places that it must be examined with care.

Under the full glare of even an equatorial sun, the temperature of the water surface rises little, for the following reasons. First, a considerable

[1] It is important, in employing graphic illustrations of this kind, that their meaning should be frequently stated in well-chosen words, and that the atmospheric conditions thus represented graphically and verbally should be clearly conceived in the mind.

share of the incident insolation is reflected away from the surface of the ocean; and this portion has no effect whatever on the temperature of the water. As it passes out through the atmosphere, a small share of it is absorbed, and the rest escapes to interplanetary space. Second, most of that which enters the water is not absorbed by the surface layer, but is transmitted to greater depths; only a little is absorbed near the surface, or in any given layer. Third, part of the absorbed insolation at the surface is expended in changing the state of some of the water from liquid to vapor or gas, and this part does not cause any rise of temperature. This is an extremely important matter, and will be referred to again and at length in Chapters VIII and IX, when treating of the moisture and the clouds of the atmosphere. It may be now simply stated that the insolation thus expended in changing the state and not in raising the temperature of the water is called the *latent heat of evaporation;* and that the evaporation of water requires a large amount of latent heat. Recalling that a unit of heat is the amount of heat needed to raise the temperature of a pound of water one degree Fahrenheit, it is found that the evaporation of a pound of water requires about a thousand units of heat; or remembering that a large calorie is the amount of heat needed to raise the temperature of a kilogram of water one degree centigrade, about 555 large calories will be needed to evaporate a kilogram of water; the precise amount varying with the temperature at which evaporation takes place. It is evident, therefore, that the evaporation of water from the ocean's surface is an effective means of retarding its rise of temperature. Fourth, the little insolation that is absorbed by the surface layer of the ocean causes only a small rise of temperature; for it is more difficult to raise the temperature of water than of any other natural substance. This matter will be referred to again when considering the case of the surface of the land. Fifth, the water of the surface of the ocean is in almost continual motion; that which is now at the surface is shortly afterwards more or less mixed with water from a greater or less depth; and that which is at one time in the torrid zone under the stronger rays of the sun, is slowly carried away by the currents, and its place is taken by other, unwarmed water. The temperature of the water at any given place is thus held down to a moderate degree by the continual removal of the warmed water and its replacement by another less warmed volume.

Nearly all these conditions operate as well in preventing a rapid fall of temperature on the ocean's surface by radiation at night as in retarding the rise of temperature by day. Being a transparent substance, and having a fairly good reflecting surface, the upper layer of water is a poor radiator. It is as difficult to cool water as it is to warm it; and the little energy lost by radiation can, therefore, have particularly little effect in lowering its temperature. As the surface layer becomes somewhat cooled, its place may then be taken by less cooled water from a depth below the surface.

The under layers of the ocean beyond the action of insolation, are even more conservative in respect to changes of temperature, either diurnal or annual, than the surface layers; and the great mass of deep water is hardly more variable in temperature than the deeper layers of the solid earth. We shall see the strong effects of the comparatively uniform temperatures of the ocean when considering the climates of different regions; those countries near the ocean, and particularly to the leeward of large oceanic areas, possess an equable temperature the year round; while those far removed from the oceans suffer from extreme ranges of temperature, more fully explained in a later section.

38. Relation of diurnal temperature range in air and water. Over the greater part of the oceans the diurnal range of temperature in the surface water is hardly one degree. The range of temperature in the open air close to the surface of the sea is two or three times as much. As this relation is the opposite of that which we shall find obtaining over the lands, it needs a brief explanation. The small amount of insolation absorbed by the lower layers of the atmosphere is all applied effectively to raising its temperature; although little is absorbed, it is so easy to warm a layer of air that the temperature is perceptibly increased. The much larger amount of insolation absorbed by the upper layer of the ocean waters is applied to various tasks; only a part of it is applied to the task of warming the water; and this task is found so difficult that the rise in the temperature at the surface is hardly noticeable. As the cooling at night is equivalent in amount to the warming by day, it follows that the lower air over the greater or oceanic surface of the globe has a stronger diurnal range of temperature than that of the surface on which it rests; but as we are dwellers on the land, this is of less importance to us than the opposite relation, namely, the stronger diurnal range of temperature in the surface layer of the ground than in the lower air, which we now have to consider.

39. Absorption and radiation by the land. The surface of the land is in many ways contrasted with that of the ocean. It is a comparatively poor reflector; but little of the insolation incident upon it is turned away to empty space. It is opaque, and all the insolation that is absorbed acts on a relatively thin surface layer. It is a solid substance, and hence cannot equalize its different temperatures by mixture, as happens in the ocean waters. It is non-volatile, and hence there is no disappearance of insolation in the form of latent heat. It is easily warmed in comparison with water, or in physical language, its specific heat is low; and hence the absorbed insolation produces a large rise of temperature. For all these reasons, the surface of the land reaches a high temperature under strong sunshine.

Conversely, it falls to a low temperature at night. It is a good radiator; all the radiant energy emitted is supplied from a thin surface layer; its specific heat is comparatively low; and hence, the active emission of radiant energy from a thin layer at the surface results in a rapid fall of temperature. Where the land is moist the changes of temperature are less than where it is dry or arid. Where it is covered by vegetation it is shielded both by day and night and its changes in temperature are greatly reduced. Moreover, in the daytime there is much evaporation from the leaves of plants; and, furthermore, there is then a certain amount of work done by insolation in separating the carbonic acid that is absorbed by the leaves into its constituent parts, as needed in the growth of the plant. Both of these processes tend to lower the temperature that would be otherwise reached.

The form of the land-surface exercises a marked control on its changes of temperature from day to night. In a valley, free, unreturned radiation to the sky is diminished by enclosure between the hillsides. On a hill-top or mountain summit, where the horizon is unbroken, radiation is most effective in reducing the temperature of the surface. This will be more fully considered in the chapter on climate.

terrestrial rays solar rays

40. Inter-radiation of air and earth. We have thus far examined the changes taking place in the different parts of the earth separately, as if they had no effect on each other; but this is not the case. The earth and the atmosphere act on each other by radiation and by conduction: both of these processes are of moment.

Just as the rays from the sun warm the surface of the earth, so the rays emitted from the surface of the earth in the daytime aid in raising the temperature of the air. Terrestrial rays, however, are weak compared to solar rays, and they are not actively absorbed by the atmosphere, but pass out through clear air with about as little loss as the solar rays suffered on entering through it. The strongest control of air temperatures by radiation from the earth will be in the lower air, near the radiating surface; over the land, whose radiation is much stronger than that of the oceans; in valleys, where the concave land surface partly encloses the air that rests on it; and when the air is somewhat dusty, so as to acquire more easily a share of the energy that passes through it.

Conversely, at night when the land-surface has fallen to a low temperature by the escape of its radiation through the atmosphere, it becomes colder than the air near it; then the air cools by moderate radiation to the cold ground. But although these processes are important, it must be observed that the changes of temperature thus produced in the air are slow and comparatively small, as well as limited for the greatest part to the lower layers of the atmosphere.

It has been supposed that the air was a better absorber of terrestrial radiation than of solar radiation; and thus the atmosphere has been compared to a trap which allowed sunshine to enter easily to the earth's surface, but prevented the free exit of radiation from the earth; water vapor, in particular, was thought to be very active in this selective process. The general temperature maintained by the atmosphere has been explained largely on these suppositions, but recent observations throw grave doubt on both of them. Clear air allows the coarse-waved radiation from the earth an easy outward passage. Water vapor is, like clear air, a poor absorber of nearly all kinds of waves. It is true that the presence of excessively fine water-particles, sufficient only to make the air faintly hazy, greatly diminishes its power of transmission, or diathermance; but water vapor, that is, water in the gaseous state, is found by experiment to be as poor an absorber as pure dry air. The temperature of the air is, therefore, now explained as a result of its own absorption and radiation, largely aided by suspended dust and by certain processes considered in the following paragraphs.

41. Conduction. It is a matter of common experience that a bar of iron heated at one end becomes heated at the other end also. This is explained by the spreading of the increased molecular agitation from the heated part to the parts less heated. Heat is thus said to flow from the hotter to the cooler parts of a body; and the passage of heat in this way is called *conduction.* It should be noticed that in this process we have nothing to do with the conversion of energy from one form or manifestation to another, as was the case both in the emission of insolation from the sun, and in its absorption on the earth. Conduction does not involve a transformation of energy, but only a distribution of energy.

Various bodies are very unlike in their ability to conduct heat. Silver and copper are good conductors; stone and water are poor conductors. From this it appears that we shall be little concerned with the downward conduction of heat from the surface of the land or water by day and in summer, or with the upward conduction by night or in winter. On the land the ordinary diurnal changes of temperature are extinguished at a depth of a few feet, and the annual changes are reduced to a very small value at a depth of twenty or thirty feet. The downward propagation of summer warmth from the surface is so slow that it is not felt at a depth of twenty-five feet until the following winter; and at that depth the annual range of temperature is reduced to somewhat less than one twentieth of its value at the surface. Fig. 3a illustrates these facts, as determined by observations at various depths at Munich, Bavaria. The time and depth at which certain temperatures occur are indicated by the curved lines, the values being given in meters and centigrade degrees.

Snow is an extremely poor conductor of heat. The surface of a sheet of snow may become extremely cold during the long clear nights of winter, when radiation goes on almost unimpeded, but this surface cooling is very slowly propagated downward, and its amount rapidly decreases underground.

The air as a whole varies in temperature so slowly from part to part that conduction within its mass has little play; except in the minute way of gaining heat from dust particles by day, and losing it to them by night, as has

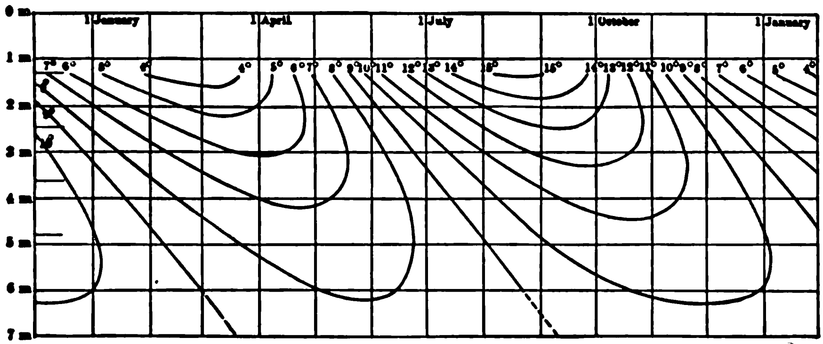


FIG. 3a.

already been referred to. But with the lower strata of air, in contact with the sea and the land, the case is different. Here it often happens that a considerable difference of temperature exists between the air and the surface on which it lies, even within a distance of a few feet; and at such times, conduction is effective. It is aided by radiation, for this, like conduction, varies directly with the contrast of temperature, and inversely with the distance between the bodies concerned: but for the moment, let us consider chiefly the case of conduction.

42. Conduction of heat between the air and the land. Consider the case of a high plateau in a northern latitude during a long winter night. Let it be far from the tempering effects of any ocean, as in the center of the great land area of Europe-Asia. The surface of the barren ground must become excessively cold. The thin, pure air above its elevated surface offers slight impediment to the escape of its heat by radiation; the dryness of such a region ensures that the sky shall be cloudless and that little or no vapor shall be condensed on the ground to retard the cooling by the liberation of latent heat; the sunshine of daytime is weak and lasts only a few hours, thus allowing the process of cooling to go on night after night with small interruption. While the ground thus cools rapidly to a low temperature, the thin, clean air high above it cools but little; but the layer of air next to the surface of the plateau, being in the neighborhood of a much colder body, loses much of its heat by

conduction to the cold ground; for while the air cannot carry much heat by conduction, the little heat that it does carry suffices very effectively to reduce the temperature of a substance so light and of so low a specific heat as air is. Supplemented by radiation, the actual cooling of the air near the ground at such a time is much greater than that of the air above it.

43. Inversions of temperature. Reference has already been made to the general decrease of temperature encountered as we ascend in the atmosphere; but in the case of the air over a dry plateau on a long winter night, the cooling of the lower layers may be so great as to reduce them to a decidedly lower temperature than that of the air at the height of several hundred feet aloft. Such a condition is known as an *inversion of temperature*. It may be illustrated in the following diagram.

Recalling the explanation given in Section 36, we have in Fig. 4 the mean vertical temperature gradient, *AB*, indicating the usual rate of decrease of temperature upwards from the plateau surface. This condition may prevail about sunset, the temperature of the air then being between the extremes of high noon and late night. When the sun's rays are no longer felt, the cooling that had begun in the afternoon is continued for a time more rapidly; and the whole mass of the atmosphere is somewhat reduced in temperature, as indicated by the horizontal lines at various altitudes. At the same time, the surface of the ground cools much more rapidly, and by midnight it may have fallen to a temperature close to Fahrenheit zero. The air near it is also greatly cooled by radiation and conduction to its cold surface, and before morning falls to a temperature, *E*, much lower than that of the air at *G*, a thousand feet above the ground. The decrease of temperature by radiation from the ground progresses rapidly at first, when it is but little cooler than the air above it; but late at night, when a strong contrast of temperature between ground and air is developed, further cooling of the ground, and thus of the air close to it, is somewhat checked by radiation from the warmer air about the height of *G*. The strong curvature of the line *EGF*, representing the peculiarly reversed vertical temperature gradient in the lower air at the late hour of greatest cold, gives clear illustration of the conditions attending such inversions of temperature as are here considered.

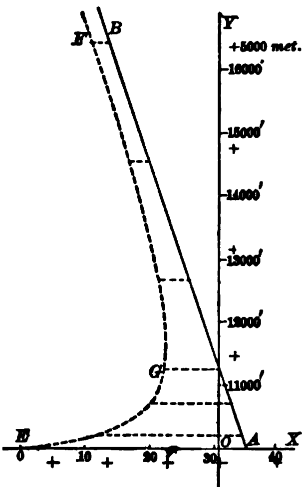


FIG. 4.

As the lower air cools, its expansive force decreases; the overlying air, no longer borne up by expansive force equal to its weight, settles down a small

distance, compressing the air beneath, and thus increasing its density and restoring its expansive force to its former equality with the weight from above. This process is not intermittent in nature, but is continually operating at every level in the atmosphere to maintain the equality between the downward weight from above and the upward expansive force from below.

Inversions of temperature are of much commoner occurrence than is generally understood. They probably occur to a greater or less degree every clear night on our dry western plains. Examples of their effects may often be seen in a small way in late spring frosts, when the lower leaves of a shrub may be nipped, while the upper branches are unharmed. In a larger way, and aided by other processes, the milder temperature of low hills than of adjacent valley bottoms at night will be explained in Section 249. It will also be shown in Section 159 that the quietness of the air at night depends largely on the occurrence of or approach to temperature inversions of the kind thus explained.

Other examples of conduction might be mentioned in the case of winds of one temperature blowing over land or water of another; but as this involves the movement of the air in large currents, it will be postponed to Section 193.

44. Convection in water. There is another process, called convection, by which unlike temperatures are partially equalized in liquids or gases. This is of great importance in the atmosphere. It may be first illustrated by a simple example in the case of water.

When a vessel of water is heated at the bottom, the warmed layer is expanded and thus made lighter than an equal volume of cooler water above it. In consequence of this unsteady arrangement, the heavier overlying water is drawn downward by gravity, displacing the bottom layer, which then rises to the surface. It is our common habit simply to say that the warmed lighter layer ascends; but it must not be forgotten that its rise is a passive process, and that the really active process is the descent of the overlying water, which is drawn down by gravity. By coloring the bottom layer, its ascent through the overlying layer may be easily perceived. If the temperature be at first uniform throughout, it will be noticed that the warmed water from the bottom is raised to the very top of the liquid, maintaining its higher temperature all the way, except for a slight loss by conduction and mixture during ascent; while all the rest of the water settles down a little distance towards the bottom. Then the new bottom layer repeats the process; and so a circulatory motion is established. This is called a convectional circulation, and by its means the entire volume of water will be warmed to almost as high a temperature as is maintained at the bottom. It depends essentially on the disturbance of a condition of rest by the introduction of a change in the temperature and a consequent change in the density of the water, which is, therefore, followed by motion under the action of gravity. After this deliberate explanation of the

convectional process, its further statement may be made more brief by speaking only of the ascent of the warm under layer, with which we are generally most concerned.

45. Conduction and convection in the atmosphere. Conduction in the atmosphere was illustrated by the cooling of the lower air at night, when it lost heat chiefly to the colder surface of the ground beneath. This change of temperature is not followed by convection, for it leaves the heaviest layer of air at the bottom, and does not give gravity any opportunity to cause motion. In the day-time, however, conduction is followed by convection, which then becomes an active process. Let us consider the case of the air over a dry plain, beneath an unclouded torrid sun. The ground warms rapidly in the morning, and soon becomes hotter than the air which rests upon it. Conduction, aided as at night by radiation, increases the temperature of the surface stratum of air. This stratum then expands, and lifts up the overlying air by a small amount, thus reversing the process of the night before. A peculiar optical effect may then be produced, which must be considered briefly.

46. Mirage.[1] As the morning advances, the lower layer of air on a level surface may become so superheated, while still lying for a time beneath the cooler heavier air, as to gain a strong vertical temperature gradient near the ground and produce the singular effect known as mirage. This is seen when the eye of the observer is a little above the surface of the superheated stratum, so as to receive the rays of light that have been reflected from it; thus frequently causing it to be mistaken for a sheet of water, with whose reflection of oblique rays from the sky to the eye we are familiar. Mirages of this kind are often observed on our barren western plains (Sect. 72).

A perfectly stagnant atmosphere might be imagined in which the alternate cooling by night and warming by day caused a corresponding rise and fall of the upper atmosphere once in twenty-four hours. In this case the work done in lifting up the upper air by day would be equal to that done in compressing the lower air at night. But such a process can hardly be supposed to proceed without interruption by currents of air of some kind.

47. Dust whirlwinds. It is not uncommon for desert mirages to disappear rather suddenly, and at the same time a local dust whirlwind springs up. This means that the superheated lower layer that has lain for a time delicately balanced under the heavier overlying air, like a layer of oil under a sheet of water, at last loses its balance and literally upsets. It then drains away upward, being urged to ascend by the descent of the heavier overlying air. The whirling of the ascending current results only because all the lines of indraft towards the point of upward escape fail to meet precisely at the

[1] A word of French origin, meaning reflection; pronounced *meerazh*.

center; they miss their aim to one side or another, and thus establish a rotary motion, which once assumed is not easily stopped. As the motion becomes brisk, dust particles are gathered up by it, vibrations are excited in its spiral currents, and the whirlwind becomes visible and audible. The dusty columns thus produced may rise to a height of a thousand or more feet, where the air currents spread out horizontally. Such whirls are not common on uneven surfaces, for there the lower air does not remain long enough close to the ground to become superheated; nor are they seen frequently on surfaces covered with vegetation, even though level; partly because such surfaces are seldom so hot as desert surfaces; partly because less dust lies upon them, by which the ascending whirls might be made visible. But the convectional ascent of the surface air in a small way is easily perceived on almost any warm, clear, quiet day by looking over the brow of a gentle rise in the ground; the air is then seen to be "unsteady," an appearance due to the passage close past one another of small currents and films of air of different temperatures, in which the rays of light are irregularly refracted. The same appearance may be seen close along side of a hot stove, and for the same reason.

It is manifest that convection must have much influence in raising the temperature of the air during the day-time; for, as long as it continues, one layer after another is brought close to the ground, where it is most effectively warmed, and whence it ascends to considerable altitudes in the atmosphere. Moreover, if no convection took place, the land-surface and the air lying close to it would become unsupportably hot under strong sunshine. In warm seasons and regions the convectional ascent of the lower air may reach a height of several miles during the hotter hours of the day, while at night the effective cooling of the air by conduction and radiation to the ground is limited to a layer a few hundred feet thick.

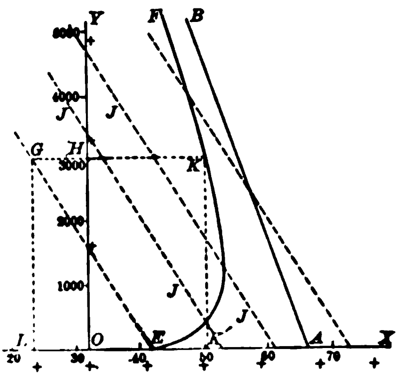
48. Difference between convection in liquids and in gases. The convectional circulation of liquids does not involve any change of temperature in the ascending and descending currents, except such as may follow mixture and conduction. With gases an important change of temperature occurs; a cooling in the ascending currents, and a warming in the descending currents of the circulation. This is entirely independent of the action of mixture and conduction. It may be briefly explained as follows.

49. Change of temperature in vertical currents. The lower air, about to ascend, has a certain temperature and a corresponding expansive force when it begins to rise. As it reaches higher levels, the pressure upon it is less; it therefore expands, pushing away the surrounding air to make room for itself, until, as a result of its expansion, its expansive force is reduced to equality with the pressure upon it. It follows, however, from experiment, as well as from the mechanical theory of heat, that in pushing away the surrounding air,

the ascending air must expend some of its energy; and this expenditure is drawn from its store of energy in the form of heat; hence the ascending air is cooled by the very processes involved in its ascent. The rate of cooling thus produced is accurately known; being ~~1.6° on the Fahrenheit scale~~ for ~~every three hundred feet~~, or 1° on the centigrade scale for 100 meters of ascent. A similar change, but of the reverse order, occurs in the descending members of the convectional circulation. As the descending air settles down, other air rolls on top of it; it is thereby compressed to a slightly greater density, and its temperature is raised. When air is thus changed in temperature, it is said to be mechanically warmed or cooled. Such changes are also called *adiabatic,* meaning thereby that they are produced without the passage of heat to or from the air.

50. Conditions of local convection in the atmosphere. The general account of convection now given makes it clear that this process cannot take place at night, when the air on the ground is colder and consequently heavier than that above it; on the other hand, convectional overturning must occur in the day-time, for then the bottom air is warmed, and may thus become light compared to that above it. But the precise amount of temperature contrast between the surface layers and the overlying air, or in other words, the precise value of the vertical temperature gradient that will allow convection, remains to be determined. A closer understanding of this problem may be gained from the following diagrams.

51. Nocturnal stability. Let *EKF*, Fig. 5, represent the value of the vertical temperature gradient in the quiet nocturnal air over a plain at a time of temperature inversion. Suppose a small volume of the surface air is raised to the altitude, *H*. As it ascends, its temperature will decrease at the adiabatic rate of 1.6° for every three hundred feet of ascent. This rate is constant at whatever temperature the ascent begins; it is indicated by the inclined line, or adiabatic gradient, *EG*.[1] When the surface air has risen to the height, *H*, its temperature will be lowered to *L*, its altitude and temperature being both indicated by the point, *G*. The temperature of the surrounding air at the height, *H*, is *K'*; hence the air that has been raised has a temperature *K'L* degrees lower than that of the air



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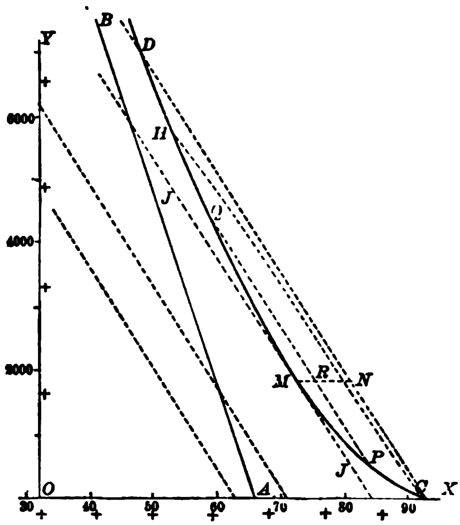
52. Diurnal instability. Consider next the conditions found at noon, when the lower air has been warmed many degrees, and the vertical temperature gradient has taken the value, *CMD*, Fig. 6. Repeat the imaginary experiment of raising a small mass of surface air to a height, *N*. From having a temperature, *C*, at the ground, it will be mechanically cooled by expansion to a temperature corresponding to the point *N*. The surrounding air at the same height has a temperature, *M*, or *MN* degrees cooler than that of the air that has been raised from the ground. The latter will therefore be lighter than the air into which it has risen, and it will continue to ascend, cooling at the adiabatic rate as it goes (no account being taken for the present of loss of heat by radiation or conduction), until it encounters air of its own temperature, as at *D*, where it will spread out laterally. Above this level it cannot rise, for at greater heights it would become colder than the surrounding air. The excess of the temperature at *C* over that at *M*, Fig. 6, is greater than occurs in nature.



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The value of the vertical temperature gradient at the time when the stability of night was changed to the instability of day should be determined. The change must have made its appearance near the ground, where the warming of the air proceeds most rapidly in the early morning. Instability occurs as soon as the line of the vertical temperature gradient is carried, in shifting from its nocturnal to its diurnal form, past parallelism with the adiabatic line. From this time on, till the warmest hour of the day is reached, the temperature of the middle air will depend chiefly on the convectional ascent of air that has been warmed close to the surface of the ground.

The same diagram may be used to determine the altitude at which the convectional ascent of the lower air will be most rapid. This will be where the

temperature of the ascending air exceeds by the greatest amount the temperature of the air through which it ascends; or at the height, *M*, where the gradient line is parallel to the adiabatic line. Moreover, all the air below this altitude is unstable, compared to the air for a certain distance above *M*. The instability of the surface air is stronger than that of any layer above it; the surface air will ascend to a greater height in the atmosphere. The air at a height, *P*, may ascend to the height, *Q*; but the air from the ground may ascend to *D*. It is manifest, however, that unless the ascending mass is of greater volume than would ordinarily be found in diurnal convection, its temperature would be reduced by mixture and conduction, as well as by expansion, during ascent; and hence it would find air of its own temperature and cease rising at some level, *H*, of less altitude than *D*. At the same time, all the air through which it rises would be warmed by the heat taken from the ascending current. Thus convection is effective in warming the lower atmosphere. The stronger the excess of temperature in the lower air, the higher it may ascend, and the more effective it will be in warming the air. Convection will therefore characterize the day-time of warm seasons and hot regions of the land, and in those seasons and regions, a considerable thickness of the lower atmosphere will be warmed by this process.

53. Explanation of convection by analogy. In any such process as this, in which motion follows a change of temperature, we find an interesting illustration of the expenditure of solar energy in the performance of work on the earth. It may be compared with the running of a clock by a weight. We wind up the weight against gravity by the expenditure of muscular energy, which is only solar energy conveniently stored for use when wanted. Gravity then pulls down the weight and sets in motion a train of wheels whose velocity is determined by the resistance of the escapement under control of the pendulum.

In an analogous manner the sun warms the lower air, which expands and raises the upper air against gravity; gravity then pulls down the upper air, and in so doing it sets certain currents in motion at a velocity determined by the resistances they encounter. Whether the motion is that of a great whirlwind or of a little filament of ascending air, it is in all cases the result of the descent somewhere else of a mass of air that has been raised against gravity by the action of insolation.

54. Local convection illustrated by clouds. A familiar effect of convection and of the adiabatic decrease of temperature that goes with it may be seen on nearly every fair summer day in the formation of clouds of greater and greater size as noon approaches, all with rather even bases at about the same altitude of a few thousand feet, and with rounded summits which may

often be seen to grow upwards, if watched carefully. Such clouds result from the convectional ascent of the lower air under the action of sunshine; for as the air rises, it gradually cools until its vapor begins to condense; then clouds begin to form; and as in a given region the lower air has a tolerably uniform temperature and moisture near the ground, the base of all the clouds formed in this way on any morning will be at about the same height. Condensation thus begun continues as long as the upward current is maintained; the convex form of the top of the ascending current is clearly shown in the rounded form of the cloud that is produced in it. Clouds of this kind, known as cumulus clouds, are common in fair summer weather over our Atlantic states; they are not so common further in the interior because there the surface air is drier, and a higher convectional ascent is necessary to produce them. They are rare on deserts, for in spite of the active convection of such regions, the surface air is so dry that ascending currents are not cooled enough by the expansion of their ascent to make them cloudy. (See also Sections 196–201.)

When the dust whirls of desert plains are carefully watched they may be seen to spread out laterally after reaching a certain height. This means that at that height their ascending air is cooled, chiefly by expansion, to the same temperature as that of the air into which it has risen; above that height it cannot go. In thunder-clouds also, which are simply examples of convection on a larger scale, a height is reached at which the temperature of the ascending current is reduced to equality with that of the air into which it ascends; at that level the cloud-bearing current spreads out laterally and produces the flat outspreading cloud-cover by which thunder-clouds may be recognized from afar, even when their thunder cannot be heard, and when their bases are below the horizon. This will be more fully considered in the chapters on clouds and local storms.

It follows from the preceding paragraphs that our atmosphere cannot have a uniform vertical distribution of temperature as long as convectional motions take place in it. However active the convection, however warm the lower air, it must cool as it rises. However long the process is continued, the upper air can never become as warm as the lower air.

55. General vertical distribution of temperature. The foregoing deliberate examination of the processes of absorption, radiation, conduction and convection should enable the reader to understand clearly the general vertical distribution of temperature in the atmosphere.

The upper air, pure and dry, free from clouds and dust, far from the surface of the earth and out of reach of ordinary convectional action, must possess a low temperature and must change its temperature slowly and by small amounts.

The lower air, containing many dusty impurities and sustaining numerous clouds, lying near the surface of the sea or land, must generally possess higher temperatures than the upper air and must generally agree closely with the temperature of the surface on which it rests. If on the ocean, its diurnal variations of temperature are small, even though a little greater than those of the ocean's surface; the temperature of the air at sea will vary chiefly with changes of the wind. If on the land, the temperature of the air varies over a strong diurnal range, and the variation thus produced is greater than that ordinarily caused by changes of the wind over a large part of the torrid land area. In the temperate zone the diurnal changes are strongest in the summer season and in clear weather, but in winter they are exceeded by the warming or cooling that accompanies the stormy shifts of the wind, as will be explained in the chapter on storms and further considered in the account of the weather.

56. Review. We are now prepared to appreciate the actual distribution of temperature over the earth in time and place. The arrangement of the atmosphere about the earth has been examined. The physical processes involved in the control of atmospheric temperatures by the sun have been carefully studied. The terrestrial sphere may be conceived as turning rapidly on its axis as it moves along its orbit, always exposing a half of its surface to the sun and thus intercepting the minutest portion of the vast shower of radiant energy emitted by that enormous globe. With the changes from day to night and from winter to summer every part of the earth is shone upon. While the parts in shadow are cooling, those under sunshine are warming; and the increase of temperature, gained chiefly at the bottom of the atmosphere, has been found to excite vertical interchanging currents by which a considerable thickness of air is warmed. The next chapter might naturally be concerned with the temperatures at different parts of the world and in different seasons of the year; but this will be postponed until another effect of insolation is examined.

CHAPTER IV.

THE COLORS OF THE SKY.

57. **The facts to be explained.** The colors of the atmosphere include those of the open sky, of the hazy air, and of the suspended clouds. The colors of clouds will be considered in a later chapter in connection with the clouds themselves. The colors of the open sky are here briefly described and explained. It is advisable to consider them under two conditions of illumination: first, when the sun stands at a considerable height above the horizon; second, when the sun is near rising or setting, either above or below the horizon.

Daytime colors. The colors of the clear sky when the sun is ten or more degrees above the horizon are for the most part shades of purer or paler blue, becoming white and glaring in the close neighborhood of the sun and turning pale or whitish towards the horizon. The clearer the weather, the purer the blue, and the less the share of white both near the sun and upon the horizon. The higher the observer rises above sea-level on mountain peaks or in balloons, the deeper the blue; the illumination of the sky is indeed then fainter, but the color produced is stronger. In the lower air the blue color fades away as haze increases, and the sky becomes whitish and more luminous; it may turn dull gray or yellowish when suspended matter is in great abundance, as in the neighborhood of forest fires.

Sunset and sunrise colors. When the sun approaches the horizon and passes below it, the intensity of sky-light decreases and the variety of color increases very greatly. As the sun sinks out of sight the most marked change from the blue of the open sky is seen in the appearance of a glowing semicircular or oval area, whose centre is somewhat above the sun and whose colors pass from a silver white through a glowing yellow to a delicate pink or purple-rose color, reaching about twenty-five degrees from the sun.

The brilliancy of the purple or rosy light is greatest when the sun is about four degrees below the horizon; its strength then decreases as the sun descends further, until when the sun is six or seven degrees below the horizon, the glow fades away. In the very clearest weather the first glow is succeeded by a second and fainter glow.

During the development of the first glow a series of horizon colors extends north and south of the point of sunset, increasing for a time in strength of coloring but at the same time decreasing in brightness. These colors are at first yellow, grading rapidly upwards through a greenish tint to the blue sky above, and fading away much more slowly along the horizon some sixty

or eighty degrees distant from the sun. As the sun descends, the yellow belt close to the horizon turns to orange and then to red ; the whole band narrowing at the same time, and fading when the depression of the sun amounts to six or seven degrees. A second but fainter series of horizon colors may accompany the second purple light. The pale western twilight that remains after the disappearance of the glows and the horizon colors, is lost when the sun is about sixteen degrees beneath the horizon; but the beginning of dawn occurs when the sun is one or more degrees further below our line of sight.

Accompanying the western colors of sunset there is a series of well-marked colors on the eastern sky. Just as the sun reaches the western horizon, the eastern horizon is marked with a pink band of color grading upwards into blue. As the sun sinks in the west, the pink band rises in the east, in the form of a long, flat arch resting on the horizon at points ninety degrees from the place of sunset. Below the pink band, which is called the twilight arch from its form and time of occurrence, there appears a belt of dull blue; in clear weather and level countries the contrast between the arch and the blue color beneath it is very distinct for some minutes after sunset: but with the rise of the arch above the eastern horizon, the sharpness of its separation from the blue belt fades away, and on reaching a height of from eight to twelve degrees it is hardly perceptible.

All of these sunset colors are seen at their best only in the clearest weather. Indeed the degree of their development may be taken as a weather prognostic, indicating the changes of a day or two to come with considerable accuracy. Turning to the opposite condition of more and more hazy or turbid atmosphere, we notice at first an increase in the strength of the yellows and reds along the horizon, and at the same time a decrease in the distinctness of the rosy glows. As the air becomes more and more turbid, the glows disappear entirely, and the horizon colors become dull, until in smoky air none of the colors appear except on the sun itself; its disc becomes orange, and finally deep crimson as it approaches the horizon; then it may even fade away before setting, leaving the western sky a dull leaden gray, without a tint of the usual sunset colors, and the eastern sky devoid of its twilight arch.

Sunrise is characterized by a very similar succession of colors, but in reverse order, and generally of somewhat fainter tints than those of sunset.

Explanation of Color in General.

58. Nature of color. Before proceeding to the special explanation of the colors of the atmosphere, a brief statement may be made of the nature and origin of colors in general. It must be remembered in the first place that the sensation of light depends upon the reception in the eye of certain undulating rays emitted by what we call luminous bodies, and transmitted by the hypo-

distance, compressing the air beneath, and thus increasing its density and restoring its expansive force to its former equality with the weight from above. This process is not intermittent in nature, but is continually operating at every level in the atmosphere to maintain the equality between the downward weight from above and the upward expansive force from below.

Inversions of temperature are of much commoner occurrence than is generally understood. They probably occur to a greater or less degree every clear night on our dry western plains. Examples of their effects may often be seen in a small way in late spring frosts, when the lower leaves of a shrub may be nipped, while the upper branches are unharmed. In a larger way, and aided by other processes, the milder temperature of low hills than of adjacent valley bottoms at night will be explained in Section 249. It will also be shown in Section 159 that the quietness of the air at night depends largely on the occurrence of or approach to temperature inversions of the kind thus explained.

Other examples of conduction might be mentioned in the case of winds of one temperature blowing over land or water of another; but as this involves the movement of the air in large currents, it will be postponed to Section 193.

44. Convection in water. There is another process, called convection, by which unlike temperatures are partially equalized in liquids or gases. This is of great importance in the atmosphere. It may be first illustrated by a simple example in the case of water.

When a vessel of water is heated at the bottom, the warmed layer is expanded and thus made lighter than an equal volume of cooler water above it. In consequence of this unsteady arrangement, the heavier overlying water is drawn downward by gravity, displacing the bottom layer, which then rises to the surface. It is our common habit simply to say that the warmed lighter layer ascends; but it must not be forgotten that its rise is a passive process, and that the really active process is the descent of the overlying water, which is drawn down by gravity. By coloring the bottom layer, its ascent through the overlying layer may be easily perceived. If the temperature be at first uniform throughout, it will be noticed that the warmed water from the bottom is raised to the very top of the liquid, maintaining its higher temperature all the way, except for a slight loss by conduction and mixture during ascent; while all the rest of the water settles down a little distance towards the bottom. Then the new bottom layer repeats the process; and so a circulatory motion is established. This is called a convectional circulation, and by its means the entire volume of water will be warmed to almost as high a temperature as is maintained at the bottom. It depends essentially on the disturbance of a condition of rest by the introduction of a change in the temperature and a consequent change in the density of the water, which is, therefore, followed by motion under the action of gravity. After this deliberate explanation of the

convectional process, its further statement may be made more brief by speaking only of the ascent of the warm under layer, with which we are generally most concerned.

45. Conduction and convection in the atmosphere. Conduction in the atmosphere was illustrated by the cooling of the lower air at night, when it lost heat chiefly to the colder surface of the ground beneath. This change of temperature is not followed by convection, for it leaves the heaviest layer of air at the bottom, and does not give gravity any opportunity to cause motion. In the day-time, however, conduction is followed by convection, which then becomes an active process. Let us consider the case of the air over a dry plain, beneath an unclouded torrid sun. The ground warms rapidly in the morning, and soon becomes hotter than the air which rests upon it. Conduction, aided as at night by radiation, increases the temperature of the surface stratum of air. This stratum then expands, and lifts up the overlying air by a small amount, thus reversing the process of the night before. A peculiar optical effect may then be produced, which must be considered briefly.

46. Mirage.[1] As the morning advances, the lower layer of air on a level surface may become so superheated, while still lying for a time beneath the cooler heavier air, as to gain a strong vertical temperature gradient near the ground and produce the singular effect known as mirage. This is seen when the eye of the observer is a little above the surface of the superheated stratum, so as to receive the rays of light that have been reflected from it; thus frequently causing it to be mistaken for a sheet of water, with whose reflection of oblique rays from the sky to the eye we are familiar. Mirages of this kind are often observed on our barren western plains (Sect. 72).

A perfectly stagnant atmosphere might be imagined in which the alternate cooling by night and warming by day caused a corresponding rise and fall of the upper atmosphere once in twenty-four hours. In this case the work done in lifting up the upper air by day would be equal to that done in compressing the lower air at night. But such a process can hardly be supposed to proceed without interruption by currents of air of some kind.

47. Dust whirlwinds. It is not uncommon for desert mirages to disappear rather suddenly, and at the same time a local dust whirlwind springs up. This means that the superheated lower layer that has lain for a time delicately balanced under the heavier overlying air, like a layer of oil under a sheet of water, at last loses its balance and literally upsets. It then drains away upward, being urged to ascend by the descent of the heavier overlying air. The whirling of the ascending current results only because all the lines of indraft towards the point of upward escape fail to meet precisely at the

[1] A word of French origin, meaning reflection; pronounced *meerazh*.

center; they miss their aim to one side or another, and thus establish a rotary motion, which once assumed is not easily stopped. As the motion becomes brisk, dust particles are gathered up by it, vibrations are excited in its spiral currents, and the whirlwind becomes visible and audible. The dusty columns thus produced may rise to a height of a thousand or more feet, where the air currents spread out horizontally. Such whirls are not common on uneven surfaces, for there the lower air does not remain long enough close to the ground to become superheated; nor are they seen frequently on surfaces covered with vegetation, even though level; partly because such surfaces are seldom so hot as desert surfaces; partly because less dust lies upon them, by which the ascending whirls might be made visible. But the convectional ascent of the surface air in a small way is easily perceived on almost any warm, clear, quiet day by looking over the brow of a gentle rise in the ground; the air is then seen to be "unsteady," an appearance due to the passage close past one another of small currents and films of air of different temperatures, in which the rays of light are irregularly refracted. The same appearance may be seen close along side of a hot stove, and for the same reason.

It is manifest that convection must have much influence in raising the temperature of the air during the day-time; for, as long as it continues, one layer after another is brought close to the ground, where it is most effectively warmed, and whence it ascends to considerable altitudes in the atmosphere. Moreover, if no convection took place, the land-surface and the air lying close to it would become unsupportably hot under strong sunshine. In warm seasons and regions the convectional ascent of the lower air may reach a height of several miles during the hotter hours of the day, while at night the effective cooling of the air by conduction and radiation to the ground is limited to a layer a few hundred feet thick.

48. Difference between convection in liquids and in gases. The convectional circulation of liquids does not involve any change of temperature in the ascending and descending currents, except such as may follow mixture and conduction. With gases an important change of temperature occurs; a cooling in the ascending currents, and a warming in the descending currents of the circulation. This is entirely independent of the action of mixture and conduction. It may be briefly explained as follows.

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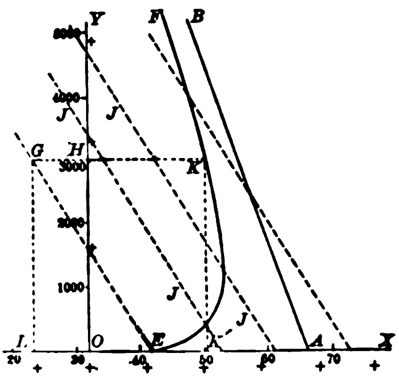

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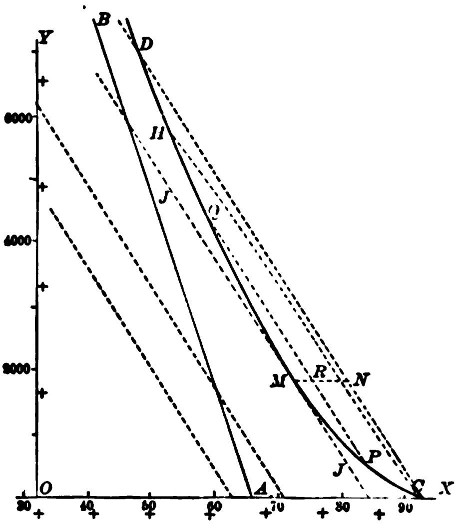

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54. Local convection illustrated by clouds. A familiar effect of convection and of the adiabatic decrease of temperature that goes with it may be seen on nearly every fair summer day in the formation of clouds of greater and greater size as noon approaches, all with rather even bases at about the same altitude of a few thousand feet, and with rounded summits which may

often be seen to grow upwards, if watched carefully. Such clouds result from the convectional ascent of the lower air under the action of sunshine; for as the air rises, it gradually cools until its vapor begins to condense; then clouds begin to form; and as in a given region the lower air has a tolerably uniform temperature and moisture near the ground, the base of all the clouds formed in this way on any morning will be at about the same height. Condensation thus begun continues as long as the upward current is maintained; the convex form of the top of the ascending current is clearly shown in the rounded form of the cloud that is produced in it. Clouds of this kind, known as cumulus clouds, are common in fair summer weather over our Atlantic states; they are not so common further in the interior because there the surface air is drier, and a higher convectional ascent is necessary to produce them. They are rare on deserts, for in spite of the active convection of such regions, the surface air is so dry that ascending currents are not cooled enough by the expansion of their ascent to make them cloudy. (See also Sections 196–201.)

When the dust whirls of desert plains are carefully watched they may be seen to spread out laterally after reaching a certain height. This means that at that height their ascending air is cooled, chiefly by expansion, to the same temperature as that of the air into which it has risen; above that height it cannot go. In thunder-clouds also, which are simply examples of convection on a larger scale, a height is reached at which the temperature of the ascending current is reduced to equality with that of the air into which it ascends; at that level the cloud-bearing current spreads out laterally and produces the flat outspreading cloud-cover by which thunder-clouds may be recognized from afar, even when their thunder cannot be heard, and when their bases are below the horizon. This will be more fully considered in the chapters on clouds and local storms.

It follows from the preceding paragraphs that our atmosphere cannot have a uniform vertical distribution of temperature as long as convectional motions take place in it. However active the convection, however warm the lower air, it must cool as it rises. However long the process is continued, the upper air can never become as warm as the lower air.

55. General vertical distribution of temperature. The foregoing deliberate examination of the processes of absorption, radiation, conduction and convection should enable the reader to understand clearly the general vertical distribution of temperature in the atmosphere.

The upper air, pure and dry, free from clouds and dust, far from the surface of the earth and out of reach of ordinary convectional action, must possess a low temperature and must change its temperature slowly and by small amounts.

conduction to the cold ground; for while the air cannot carry much heat by conduction, the little heat that it does carry suffices very effectively to reduce the temperature of a substance so light and of so low a specific heat as air is. Supplemented by radiation, the actual cooling of the air near the ground at such a time is much greater than that of the air above it.

43. Inversions of temperature. Reference has already been made to the general decrease of temperature encountered as we ascend in the atmosphere; but in the case of the air over a dry plateau on a long winter night, the cooling of the lower layers may be so great as to reduce them to a decidedly lower temperature than that of the air at the height of several hundred feet aloft. Such a condition is known as an *inversion of temperature.* It may be illustrated in the following diagram.

Recalling the explanation given in Section 36, we have in Fig. 4 the mean vertical temperature gradient, *AB*, indicating the usual rate of decrease of temperature upwards from the plateau surface. This condition may prevail about sunset, the temperature of the air then being between the extremes of high noon and late night. When the sun's rays are no longer felt, the cooling that had begun in the afternoon is continued for a time more rapidly; and the whole mass of the atmosphere is somewhat reduced in temperature, as indicated by the horizontal lines at various altitudes. At the same time, the surface of the ground cools much more rapidly, and by midnight it may have fallen to a temperature close to Fahrenheit zero. The air near it is also greatly cooled by radiation and conduction to its cold surface, and before morning falls to a temperature, *E*, much lower than that of the air at *G*, a thousand feet above the ground. The decrease of temperature by radiation from the ground progresses rapidly at first, when it is but little cooler than the air above it; but late at night, when a strong contrast of temperature between ground and air is developed, further cooling of the ground, and thus of the air close to it, is somewhat checked by radiation from the warmer air about the height of *G*. The strong curvature of the line *EGF*, representing the peculiarly reversed vertical temperature gradient in the lower air at the late hour of greatest cold, gives clear illustration of the conditions attending such inversions of temperature as are here considered.

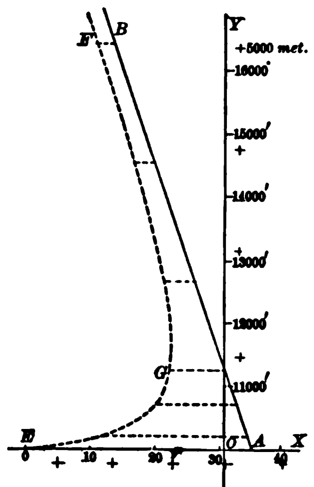

Fig. 4.

As the lower air cools, its expansive force decreases; the overlying air, no longer borne up by expansive force equal to its weight, settles down a small

distance, compressing the air beneath, and thus increasing its density and restoring its expansive force to its former equality with the weight from above. This process is not intermittent in nature, but is continually operating at every level in the atmosphere to maintain the equality between the downward weight from above and the upward expansive force from below.

Inversions of temperature are of much commoner occurrence than is generally understood. They probably occur to a greater or less degree every clear night on our dry western plains. Examples of their effects may often be seen in a small way in late spring frosts, when the lower leaves of a shrub may be nipped, while the upper branches are unharmed. In a larger way, and aided by other processes, the milder temperature of low hills than of adjacent valley bottoms at night will be explained in Section 249. It will also be shown in Section 159 that the quietness of the air at night depends largely on the occurrence of or approach to temperature inversions of the kind thus explained.

Other examples of conduction might be mentioned in the case of winds of one temperature blowing over land or water of another; but as this involves the movement of the air in large currents, it will be postponed to Section 193.

44. Convection in water. There is another process, called convection, by which unlike temperatures are partially equalized in liquids or gases. This is of great importance in the atmosphere. It may be first illustrated by a simple example in the case of water.

When a vessel of water is heated at the bottom, the warmed layer is expanded and thus made lighter than an equal volume of cooler water above it. In consequence of this unsteady arrangement, the heavier overlying water is drawn downward by gravity, displacing the bottom layer, which then rises to the surface. It is our common habit simply to say that the warmed lighter layer ascends; but it must not be forgotten that its rise is a passive process, and that the really active process is the descent of the overlying water, which is drawn down by gravity. By coloring the bottom layer, its ascent through the overlying layer may be easily perceived. If the temperature be at first uniform throughout, it will be noticed that the warmed water from the bottom is raised to the very top of the liquid, maintaining its higher temperature all the way, except for a slight loss by conduction and mixture during ascent; while all the rest of the water settles down a little distance towards the bottom. Then the new bottom layer repeats the process; and so a circulatory motion is established. This is called a convectional circulation, and by its means the entire volume of water will be warmed to almost as high a temperature as is maintained at the bottom. It depends essentially on the disturbance of a condition of rest by the introduction of a change in the temperature and a consequent change in the density of the water, which is, therefore, followed by motion under the action of gravity. After this deliberate explanation of the

convectional process, its further statement may be made more brief by speaking only of the ascent of the warm under layer, with which we are generally most concerned.

45. Conduction and convection in the atmosphere. Conduction in the atmosphere was illustrated by the cooling of the lower air at night, when it lost heat chiefly to the colder surface of the ground beneath. This change of temperature is not followed by convection, for it leaves the heaviest layer of air at the bottom, and does not give gravity any opportunity to cause motion. In the day-time, however, conduction is followed by convection, which then becomes an active process. Let us consider the case of the air over a dry plain, beneath an unclouded torrid sun. The ground warms rapidly in the morning, and soon becomes hotter than the air which rests upon it. Conduction, aided as at night by radiation, increases the temperature of the surface stratum of air. This stratum then expands, and lifts up the overlying air by a small amount, thus reversing the process of the night before. A peculiar optical effect may then be produced, which must be considered briefly.

46. Mirage.[1] As the morning advances, the lower layer of air on a level surface may become so superheated, while still lying for a time beneath the cooler heavier air, as to gain a strong vertical temperature gradient near the ground and produce the singular effect known as mirage. This is seen when the eye of the observer is a little above the surface of the superheated stratum, so as to receive the rays of light that have been reflected from it; thus frequently causing it to be mistaken for a sheet of water, with whose reflection of oblique rays from the sky to the eye we are familiar. Mirages of this kind are often observed on our barren western plains (Sect. 72).

A perfectly stagnant atmosphere might be imagined in which the alternate cooling by night and warming by day caused a corresponding rise and fall of the upper atmosphere once in twenty-four hours. In this case the work done in lifting up the upper air by day would be equal to that done in compressing the lower air at night. But such a process can hardly be supposed to proceed without interruption by currents of air of some kind.

47. Dust whirlwinds. It is not uncommon for desert mirages to disappear rather suddenly, and at the same time a local dust whirlwind springs up. This means that the superheated lower layer that has lain for a time delicately balanced under the heavier overlying air, like a layer of oil under a sheet of water, at last loses its balance and literally upsets. It then drains away upward, being urged to ascend by the descent of the heavier overlying air. The whirling of the ascending current results only because all the lines of indraft towards the point of upward escape fail to meet precisely at the

[1] A word of French origin, meaning reflection; pronounced *meerazh.*

center; they miss their aim to one side or another, and thus establish a rotary motion, which once assumed is not easily stopped. As the motion becomes brisk, dust particles are gathered up by it, vibrations are excited in its spiral currents, and the whirlwind becomes visible and audible. The dusty columns thus produced may rise to a height of a thousand or more feet, where the air currents spread out horizontally. Such whirls are not common on uneven surfaces, for there the lower air does not remain long enough close to the ground to become superheated; nor are they seen frequently on surfaces covered with vegetation, even though level; partly because such surfaces are seldom so hot as desert surfaces; partly because less dust lies upon them, by which the ascending whirls might be made visible. But the convectional ascent of the surface air in a small way is easily perceived on almost any warm, clear, quiet day by looking over the brow of a gentle rise in the ground; the air is then seen to be "unsteady," an appearance due to the passage close past one another of small currents and films of air of different temperatures, in which the rays of light are irregularly refracted. The same appearance may be seen close along side of a hot stove, and for the same reason.

It is manifest that convection must have much influence in raising the temperature of the air during the day-time; for, as long as it continues, one layer after another is brought close to the ground, where it is most effectively warmed, and whence it ascends to considerable altitudes in the atmosphere. Moreover, if no convection took place, the land-surface and the air lying close to it would become unsupportably hot under strong sunshine. In warm seasons and regions the convectional ascent of the lower air may reach a height of several miles during the hotter hours of the day, while at night the effective cooling of the air by conduction and radiation to the ground is limited to a layer a few hundred feet thick.

48. Difference between convection in liquids and in gases. The convectional circulation of liquids does not involve any change of temperature in the ascending and descending currents, except such as may follow mixture and conduction. With gases an important change of temperature occurs; a cooling in the ascending currents, and a warming in the descending currents of the circulation. This is entirely independent of the action of mixture and conduction. It may be briefly explained as follows.

49. Change of temperature in vertical currents. The lower air, about to ascend, has a certain temperature and a corresponding expansive force when it begins to rise. As it reaches higher levels, the pressure upon it is less; it therefore expands, pushing away the surrounding air to make room for itself, until, as a result of its expansion, its expansive force is reduced to equality with the pressure upon it. It follows, however, from experiment, as well as from the mechanical theory of heat, that in pushing away the surrounding air,

the ascending air must expend some of its energy; and this expenditure is drawn from its store of energy in the form of heat; hence the ascending air is cooled by the very processes involved in its ascent. The rate of cooling thus produced is accurately known; being ~~1.6° on the Fahrenheit scale~~ for ~~every three hundred feet~~, or 1° on the centigrade scale for 100 meters of ascent. A similar change, but of the reverse order, occurs in the descending members of the convectional circulation. As the descending air settles down, other air rolls on top of it; it is thereby compressed to a slightly greater density, and its temperature is raised. When air is thus changed in temperature, it is said to be mechanically warmed or cooled. Such changes are also called *adiabatic*, meaning thereby that they are produced without the passage of heat to or from the air.

50. Conditions of local convection in the atmosphere. The general account of convection now given makes it clear that this process cannot take place at night, when the air on the ground is colder and consequently heavier than that above it; on the other hand, convectional overturning must occur in the day-time, for then the bottom air is warmed, and may thus become light compared to that above it. But the precise amount of temperature contrast between the surface layers and the overlying air, or in other words, the precise value of the vertical temperature gradient that will allow convection, remains to be determined. A closer understanding of this problem may be gained from the following diagrams.

51. Nocturnal stability. Let *EKF*, Fig. 5, represent the value of the vertical temperature gradient in the quiet nocturnal air over a plain at a time of temperature inversion. Suppose a small volume of the surface air is raised to the altitude, *H*. As it ascends, its temperature will decrease at the adiabatic rate of 1.6° for every three hundred feet of ascent. This rate is constant at whatever temperature the ascent begins; it is indicated by the inclined line, or adiabatic gradient, *EG*.[1] When the surface air has risen to the height, *H*, its temperature will be lowered to *L*, its altitude and temperature being both indicated by the point, *G*. The temperature of the surrounding air at the height, *H*, is *K'*; hence the air

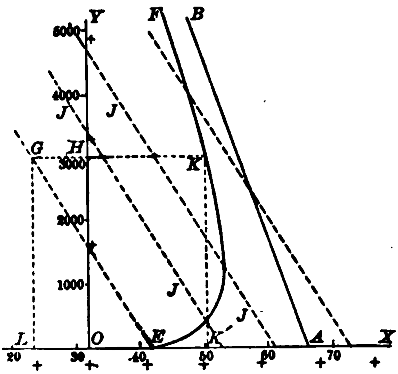

Fig. 5.

that has been raised has a temperature *K'L* degrees lower than that of the air

[1] In all diagrams of this kind the adiabatic rate of cooling will be indicated by straight broken lines.

that it has risen into. It will therefore be much heavier than the surrounding air, and consequently, if no longer sustained, it will sink down to the ground before finding any air of its own temperature. It must be concluded from this that the lower air on plains during clear, quiet nights is not disposed to move; and that if disturbed, it will tend to return to the position that it had before the disturbance. The air is then said to be in a stable equilibrium.

52. Diurnal instability. Consider next the conditions found at noon, when the lower air has been warmed many degrees, and the vertical temperature gradient has taken the value, *CMD*, Fig. 6. Repeat the imaginary experiment of raising a small mass of surface air to a height, *N*. From having a temperature, *C*, at the ground, it will be mechanically cooled by expansion to a temperature corresponding to the point *N*. The surrounding air at the same height has a temperature, *M*, or *MN* degrees cooler than that of the air that has been raised from the ground. The latter will therefore be lighter than the air into which it has risen, and it will continue to ascend, cooling at the adiabatic rate as it goes (no account being taken for the present of loss of heat by radiation or conduction), until it encounters air of its own temperature, as at *D*, where it will spread out laterally. Above this level it cannot rise, for at greater heights it would become colder than the surrounding air. The excess of the temperature at *C* over that at *M*, Fig. 6, is greater than occurs in nature.

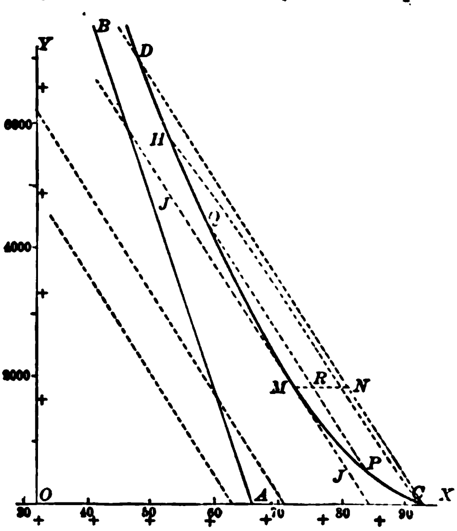

Fig. 6.

The value of the vertical temperature gradient at the time when the stability of night was changed to the instability of day should be determined. The change must have made its appearance near the ground, where the warming of the air proceeds most rapidly in the early morning. Instability occurs as soon as the line of the vertical temperature gradient is carried, in shifting from its nocturnal to its diurnal form, past parallelism with the adiabatic line. From this time on, till the warmest hour of the day is reached, the temperature of the middle air will depend chiefly on the convectional ascent of air that has been warmed close to the surface of the ground.

The same diagram may be used to determine the altitude at which the convectional ascent of the lower air will be most rapid. This will be where the

temperature of the ascending air exceeds by the greatest amount the temperature of the air through which it ascends; or at the height, M, where the gradient line is parallel to the adiabatic line. Moreover, all the air below this altitude is unstable, compared to the air for a certain distance above M. The instability of the surface air is stronger than that of any layer above it; the surface air will ascend to a greater height in the atmosphere. The air at a height, P, may ascend to the height, Q; but the air from the ground may ascend to D. It is manifest, however, that unless the ascending mass is of greater volume than would ordinarily be found in diurnal convection, its temperature would be reduced by mixture and conduction, as well as by expansion, during ascent; and hence it would find air of its own temperature and cease rising at some level, H, of less altitude than D. At the same time, all the air through which it rises would be warmed by the heat taken from the ascending current. Thus convection is effective in warming the lower atmosphere. The stronger the excess of temperature in the lower air, the higher it may ascend, and the more effective it will be in warming the air. Convection will therefore characterize the day-time of warm seasons and hot regions of the land, and in those seasons and regions, a considerable thickness of the lower atmosphere will be warmed by this process.

53. Explanation of convection by analogy. In any such process as this, in which motion follows a change of temperature, we find an interesting illustration of the expenditure of solar energy in the performance of work on the earth. It may be compared with the running of a clock by a weight. We wind up the weight against gravity by the expenditure of muscular energy, which is only solar energy conveniently stored for use when wanted. Gravity then pulls down the weight and sets in motion a train of wheels whose velocity is determined by the resistance of the escapement under control of the pendulum.

In an analogous manner the sun warms the lower air, which expands and raises the upper air against gravity; gravity then pulls down the upper air, and in so doing it sets certain currents in motion at a velocity determined by the resistances they encounter. Whether the motion is that of a great whirlwind or of a little filament of ascending air, it is in all cases the result of the descent somewhere else of a mass of air that has been raised against gravity by the action of insolation.

54. Local convection illustrated by clouds. A familiar effect of convection and of the adiabatic decrease of temperature that goes with it may be seen on nearly every fair summer day in the formation of clouds of greater and greater size as noon approaches, all with rather even bases at about the same altitude of a few thousand feet, and with rounded summits which may

often be seen to grow upwards, if watched carefully. Such clouds result from the convectional ascent of the lower air under the action of sunshine, for as the air rises it gradually cools until its vapor begins to condense; then clouds begin to form; and as in a given region the lower air has a tolerably uniform temperature and moisture near the ground, the base of all the clouds formed in this way on any morning will be at about the same height. Condensation thus begun continues as long as the upward current is maintained; the convex form of the top of the ascending current is clearly shown in the rounded form of the cloud that is produced in it. Clouds of this kind, known as cumulus clouds, are common in fair summer weather over our Atlantic states; they are not so common further in the interior because there the surface air is drier, and a higher convectional ascent is necessary to produce them. They are rare on deserts, for in spite of the active convection of such regions, the surface air is so dry that ascending currents are not cooled enough by the expansion of their ascent to make them cloudy. (See also Sections 196–201.)

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It follows from the preceding paragraphs that our atmosphere cannot have a uniform vertical distribution of temperature as long as convectional motions take place in it. However active the convection, however warm the lower air, it must cool as it rises. However long the process is continued, the upper air can never become as warm as the lower air.

55. General vertical distribution of temperature. The foregoing deliberate examination of the processes of absorption, radiation, conduction and convection should enable the reader to understand clearly the general vertical distribution of temperature in the atmosphere.

The upper air, pure and dry, free from clouds and dust, far from the surface of the earth and out of reach of ordinary convectional currents, must possess a low temperature and must change its temperature slowly and by small amounts.

The lower air, containing many dusty impurities and sustaining numerous clouds, lying near the surface of the sea or land, must generally possess higher temperatures than the upper air and must generally agree closely with the temperature of the surface on which it rests. If on the ocean, its diurnal variations of temperature are small, even though a little greater than those of the ocean's surface; the temperature of the air at sea will vary chiefly with changes of the wind. If on the land, the temperature of the air varies over a strong diurnal range, and the variation thus produced is greater than that ordinarily caused by changes of the wind over a large part of the torrid land area. In the temperate zone the diurnal changes are strongest in the summer season and in clear weather, but in winter they are exceeded by the warming or cooling that accompanies the stormy shifts of the wind, as will be explained in the chapter on storms and further considered in the account of the weather.

56. Review. We are now prepared to appreciate the actual distribution of temperature over the earth in time and place. The arrangement of the atmosphere about the earth has been examined. The physical processes involved in the control of atmospheric temperatures by the sun have been carefully studied. The terrestrial sphere may be conceived as turning rapidly on its axis as it moves along its orbit, always exposing a half of its surface to the sun and thus intercepting the minutest portion of the vast shower of radiant energy emitted by that enormous globe. With the changes from day to night and from winter to summer every part of the earth is shone upon. While the parts in shadow are cooling, those under sunshine are warming; and the increase of temperature, gained chiefly at the bottom of the atmosphere, has been found to excite vertical interchanging currents by which a considerable thickness of air is warmed. The next chapter might naturally be concerned with the temperatures at different parts of the world and in different seasons of the year; but this will be postponed until another effect of insolation is examined.

CHAPTER IV.

THE COLORS OF THE SKY.

57. The facts to be explained. The colors of the atmosphere [illegible] those of the open sky, of the hazy air, and of the [illegible]. The colors of clouds will be considered in a later chapter [illegible] with the clouds themselves. The colors of the open sky are [illegible] explained. It is advisable to consider them under two [illegible] tion: first, when the sun stands at a considerable [illegible] second, when the sun is near rising or setting, [illegible] horizon.

Daytime colors. The colors of the clear sky [illegible] degrees above the horizon are for the most part [illegible] becoming white and glaring in the close neighborhood of [illegible] pale or whitish towards the horizon. The [illegible] blue, and the less the share of white [illegible]. The higher the observer rises above [illegible] the deeper the blue; the illumination of the [illegible] the color produced is stronger. In the lower air [illegible] haze increases, and the sky becomes [illegible] dull gray or yellowish when suspended matter [illegible] neighborhood of forest fires.

Sunset and sunrise colors. When the sun [illegible] horizon and passes below it, the intensity of sky-light [illegible] increases very greatly. As the sun [illegible] from the blue of the open sky is [illegible] circular or oval area, whose centre is [illegible] pass from a silver white through a [illegible] rose color, reaching about twenty-five [illegible].

The brilliancy of the purple or rosy [illegible] about four degrees below the horizon; [illegible] as the sun descends further, until when the sun [illegible] the glow fades away. In the very [illegible] by a second and fainter glow.

During the development of the first glow a [illegible] extends north and south of the point of sunset, [illegible] strength of coloring but at the same time [illegible]. These colors are at first yellow, grading rapidly upwards through a greenish tint to the blue sky above, and fading away much more slowly along the horizon some sixty

or eighty degrees distant from the sun. As the sun descends, the yellow belt close to the horizon turns to orange and then to red ; the whole band narrowing at the same time, and fading when the depression of the sun amounts to six or seven degrees. A second but fainter series of horizon colors may accompany the second purple light. The pale western twilight that remains after the disappearance of the glows and the horizon colors, is lost when the sun is about sixteen degrees beneath the horizon ; but the beginning of dawn occurs when the sun is one or more degrees further below our line of sight.

Accompanying the western colors of sunset there is a series of well-marked colors on the eastern sky. Just as the sun reaches the western horizon, the eastern horizon is marked with a pink band of color grading upwards into blue. As the sun sinks in the west, the pink band rises in the east, in the form of a long, flat arch resting on the horizon at points ninety degrees from the place of sunset. Below the pink band, which is called the twilight arch from its form and time of occurrence, there appears a belt of dull blue; in clear weather and level countries the contrast between the arch and the blue color beneath it is very distinct for some minutes after sunset: but with the rise of the arch above the eastern horizon, the sharpness of its separation from the blue belt fades away, and on reaching a height of from eight to twelve degrees it is hardly perceptible.

All of these sunset colors are seen at their best only in the clearest weather. Indeed the degree of their development may be taken as a weather prognostic, indicating the changes of a day or two to come with considerable accuracy. Turning to the opposite condition of more and more hazy or turbid atmosphere, we notice at first an increase in the strength of the yellows and reds along the horizon, and at the same time a decrease in the distinctness of the rosy glows. As the air becomes more and more turbid, the glows disappear entirely, and the horizon colors become dull, until in smoky air none of the colors appear except on the sun itself; its disc becomes orange, and finally deep crimson as it approaches the horizon; then it may even fade away before setting, leaving the western sky a dull leaden gray, without a tint of the usual sunset colors, and the eastern sky devoid of its twilight arch.

Sunrise is characterized by a very similar succession of colors, but in reverse order, and generally of somewhat fainter tints than those of sunset.

Explanation of Color in General.

58. Nature of color. Before proceeding to the special explanation of the colors of the atmosphere, a brief statement may be made of the nature and origin of colors in general. It must be remembered in the first place that the sensation of light depends upon the reception in the eye of certain undulating rays emitted by what we call luminous bodies, and transmitted by the hypo-

thetical ether, already explained in Section 24. Moreover, luminous bodies send out rays of a great variety of wave-lengths, many of which our eyes cannot perceive, perhaps because the media of the eye are not transparent to them. Only the rays whose wave-length is between 0.00036 and 0.00075 mm. can be seen. It is possible that "seeing light" is simply the sensation of heat produced in the optic nerves by the absorption of the rays that reach the retina. The greater the amplitude of the waves, the more intense the light.

Color is determined by the proportion of rays of different wave-length. When the intensity of the various rays exists in the proportions occurring in ordinary sunlight, we get no sensation of color, and the light is then called white. If any of the rays are unduly intense, or if others are unduly weakened, the light is colored. A red light, for example, is one in which the coarser rays are more intense; a blue light, one in which the finer waves are more intense. But in all natural colors there are rays of a great variety of wave-length, and the color that we perceive is determined only by the action of the more intense rays. This is easily shown by looking at a colored object through a prism. The green of foliage, for example, is thus found to be merely green in excess; nearly all other colors of the spectrum being perceptible in it. So the blue of the sky or the red of sunset contains an almost full series of other colors, but the rays which determine the color are of the greatest intensity.

If sunlight in the ordinary proportions gives the sensation of white light, we must inquire into the processes by which its normal composition is so changed as to give the various colors of the sky at one time and another.

59. Selective absorption and diffuse reflection. Many solid opaque substances absorb rays of one wave-length better than those of another. The rays incident on such bodies are thus divided into two classes; the one absorbed, and the other irregularly turned back or diffusely reflected.[1] It has already been explained in the chapter on the temperature of the atmosphere that the absorbed rays raise the temperature of the absorber and increase the intensity of the radiation emitted by it; but at ordinary temperatures the rays thus emitted are of great wave-length, quite imperceptible by the eye. On the other hand, the diffusely reflected rays depart unchanged in wave-length; but the proportion of various wave-lengths in the reflected light is greatly altered from the normal proportions in the incident light. The illuminated object then appears to have a color corresponding to that of the rays that are in excess. Thus the green of foliage is produced; not that all colors but green are absorbed and green alone is reflected, but that green is

[1] It is probable that diffraction also takes place to a large extent on the rough surface of ordinary substances; but the whole process is commonly included under the term, "diffuse reflection."

less absorbed and more turned aside than the other colors. This process is more important than any other in determining the color of objects on the earth's surface; but it is not known to have application in causing the colors of the atmosphere.

60. Selective absorption and transmission. Transparent substances, whether solid, liquid or gaseous, are sometimes colorless, sometimes colored. Colorless objects, such as glass or water, permit the nearly free passage of the optical rays, although they may absorb rays of other wave-lengths. Colored transparent substances exercise a selective absorption on certain of the optical rays, allowing the others to pass, and thus determine their color; thus the yellow of amber, the tints of various inks and the green of chlorine gas are produced. It is possible that this process has a share in determining some of the atmospheric colors; but it does not control them, as will appear further on.

61. Selective scattering. When a transparent substance, either solid, liquid or gaseous, contains suspended in it a great number of excessively fine particles whose dimensions are smaller than the wave-lengths of light, the rays that would otherwise be allowed to pass unobstructed are scattered or diffracted in all directions on every particle; but the finer waved rays are more effectively turned from their path than the coarser waved rays. The light that passes through such a turbid medium in the direction of the original rays therefore becomes more or less yellow or red; while that which departs laterally has finer waved rays in excess, and appears blue. This may be illustrated by a simple experiment with a flask of soapy water; on looking through it towards a source of white light, the liquid appears somewhat yellowish or orange; looking across the direction of the illuminating rays, the same liquid appears to be bluish. A column of smoke may also appear of different colors, according to its illumination. If looked at against the sky, it seems to be brownish yellow; if viewed against a dark background of heavily shaded trees, it appears blue. This process of selective scattering is of much importance in explaining the colors of the sky.

62. Diffraction and interference. The particles in a turbid medium may be of larger dimensions than the wave lengths of light. Then the rays that are scattered or diffracted from the opposite sides of a single particle may "interfere" and extinguish each other. This process cannot be explained here; but its effects may be examined by looking at a source of white light through a glass plate on which lycopodium powder is scattered. The light will appear to be surrounded by concentric rings of prismatic colors, with blue on the inside and red on the outside. If the diffracting particles are numerous, the rings will be bright. If the particles are all of one size, the rings

will be sharply defined with distinct colors. Large particles produce rings of small diameter; small particles produce large rings. If the particles are of many sizes, the colors of the large and small rings overlap and blend into a disc of white light, 'brighest close to the center; the disc may be bordered with a reddish tinge, when the marginal color of the outer ring is not wholly lost. Such a disc may be called a diffraction glow. This process is important in explaining certain sunset colors, as well as in accounting for the coronas or colored rings around the sun or moon in thin clouds.

63. Refraction. When a beam of light passes from a rarer to a denser medium, it is bent towards the vertical to the surface separating the two media; and the finer waved rays are bent more than those of greater wave lengths. It is thus that white sunlight is broken or refracted into the colors of the spectrum when passing through a transparent prism. We shall find application of this process chiefly in accounting for halos and rainbows in a later chapter.

EXPLANATION OF THE COLORS OF THE SKY.

64. The dust of the atmosphere. The general blue color of the sky is best accounted for by the process of selective scattering, as explained by Lord Rayleigh. As this depends on the presence of innumerable sub-microscopic, non-gaseous particles in the atmosphere, a paragraph may be given to their probable origin and constitution.

The atmosphere is known to contain a vast number of minute particles, solid for the most part, and commonly named *dust*. The coarser particles will settle from a body of air if it is allowed to rest quietly, and in speaking of dust we commonly refer only to such particles as can easily be collected from the air when it is still. But there is very good reason to think that the impurities of the atmosphere include myriads of particles vastly finer than those to which the name, dust, is ordinarily applied, and in speaking of atmospheric dust in this chapter, all particles from the coarsest to the finest will be included.

The atmosphere receives its dust chiefly from the earth. It is carried up from the lands by the wind; it is blown out of active volcanoes; some of it comes from salt in the ocean, remaining in the air when spray from the waves is evaporated. An undetermined share must come from the combustion of meteors high above sea-level. Water vapor, condensed into the minutest drops of water or crystals of ice, may provide much of the so-called dust. The coarser dust for the most part being received from the land surfaces at the bottom of the atmosphere, it is natural that its greatest amount should be found in the lower layers of air over the continents; but it is borne so easily

by the winds that it is carried far and wide over the earth. Vessels in the Atlantic, west of the Sahara, may have their sails reddened by falling dust that has been carried out from the desert by the trade winds. As we ascend above sea-level, even in regions reputed to have a clear atmosphere, as in Italy or on the Azores, the lower strata seem like a dusty ocean above which the clearer air of the higher regions floats. Yet it is conceived that even the upper air contains innumerable particles of a size so small that in spite of their distribution through the great volume of atmosphere, their total quantity is not great, and their falling towards the earth is prevented by the faintest ascensional current. Although below microscopic sight, they must not be confused with the molecules of the atmospheric gases, which are to the best of our knowledge of a far greater degree of minuteness. The atmosphere must therefore be regarded, even when apparently clearest, as a slightly turbid medium.

65. The blue of the sky: selective scattering. When a beam of white light passes through the turbid atmosphere, the rays laterally scattered in all directions from the path of the beam will contain a greater share of blue than of red light. The coarser the particles, the greater the intensity of the scattered rays, and the more uniform the proportion of the rays of different wave-lengths. The finer the particles, the fainter the scattered rays, but the greater the excess of blue in the rays turned aside from the original beam.

Now in looking up into the sky, away from the sun, the light that comes to our eyes is that which has been scattered from many solar rays as they encounter the myriads of suspended particles which for the moment happen to be in our line of sight; and as these particles are more effective in turning aside the fine-waved rays than the coarser ones, the eye receives them in excess, and the sky appears blue.[1] It is manifest that the term, reflection, should not be used in describing this process.

Many relatively coarse particles in the lower air add reflected white light to the blue color produced by the smaller particles, and thus increase the illumination and whiteness of the sky. When the air is hazy, the larger particles predominate, and the blue is almost lost in a whitish glare; not from the cessation of the process by which the blue is made, but simply by the addition of a greater and greater quantity of white light from the more abundant and larger particles. When the air is very clear, the action of the

[1] *Polarization.* After the scattering of rays on fine particles, their waves vibrate more or less perfectly in a single plane, instead of in all directions transverse to the ray. The ray is then said to be *polarized.* Special instruments are devised to detect this peculiarity; and from these it is found that the light of the sky is polarized to a greater or less degree; the most complete polarization occurring at an angular distance of 90° from the sun. This is found to be a necessary consequence of the scattering of sunlight on extremely fine particles, as has been shown by Lord Rayleigh, by whom this theory of sky color was advanced.

finer particles, especially of those always present in the upper air, produces the deep blue of the celestial vault.

The greater brilliancy and whiteness of the sky near the sun finds its explanation in the statement already made concerning the greater intensity of diffraction from fine particles nearly in line with the original ray. As we look closer and closer towards the sun, our line of sight is more nearly in the line of the direct rays, and the intensity of the light is correspondingly increased. The composition of slightly diffracted rays is so nearly the same as that of the direct rays that they appear like normal sunlight, or white.

The sky near the horizon becomes whiter and paler than at greater altitudes. This is because a line of sight passes through a much greater distance of lower dusty air when we look near the horizon than when we look towards the zenith; the numerous coarse motes thus encountered turn white light to the eye and overpower the blue that comes with it.

66. The color of the sun. The reader may now naturally inquire why the direct rays of the sun appear white. According to the explanation just given, the rays that come from the sun directly to us have lost a greater share of blue than of red rays, and the sun should therefore appear of a reddened or at least of an orange tint. This reasoning is correct, and the best answer to it is the one suggested by Langley. If we could see the sun from outside of our atmosphere, it would probably be a blue sun; it appears white only after an excess of blue in the original sunbeam has been turned away on its passage to us through the dusty air. In this respect the sun is not a unique body; blue stars are known in various parts of the sky, and these are evidently even more blue than our sun; otherwise their light also would be white when it came down to us.

67. Deep blue sky seen from mountains. The greater purity and fainter illumination of the blue of the sky when seen from lofty mountain tops is due to the absence in the upper air of those larger dust motes so common at lower levels. The larger motes turn to us waves of all kinds, diluting the blue of the sky by the addition of white light to the observer at sea-level. When we rise above the level at which the coarser motes are common, their action weakens; the sky is less illuminated, but of a deeper and purer blue color; and at heights of fifteen or twenty thousand feet observers describe it as of a deep indigo color, extremely dark compared to the well illuminated sky to which we are accustomed.

68. Sunset and sunrise horizon colors. The colors of evening and morning are essentially the same, but they are exhibited in reverse order. Those of sunset will be here explained, and the tints of sunrise will be referred to only when special mention of them is needed.

The sunset colors have already been divided into two series; one arranged along the horizon, the other disposed in a circular segment or glow with the sun near the center. The former will be first considered.

The solar disc itself is yellow or red as it sets, because then its direct rays have traversed so great a thickness of air that the blues are greatly diminished by selective scattering, leaving the others in excess. As the sun nears the horizon, the lower western sky, where the color was whitish at noon, becomes yellowish, by reason of the scattering away of the blue rays. When the sun is below the horizon, the yellows and reds that fringe the sky-line north and south from the point of setting, are for the most part due in the same way to the effective scattering of the finer rays in passing through the great thickness of air that they then must traverse. The beams of light that come through the greatest thickness of air, close to the horizon, are the most strongly reddened. The beams from a belt a few degrees above the horizon have passed through a less measure of atmosphere and by a less direct course, and are therefore orange or yellow, instead of red. None of the horizon colors, however, are direct rays; they all suffer more or less bending on their way, especially those that come from points some distance north or south of the point of sunset. The greater their bending, the less intense their red color. Much of the bending may be done by the larger dust motes, which send white light in the day-time; at sunset these simply send along the light that falls on them without affecting its color. They thus serve to extend the reds and yellows along the western horizon, but not to alter the color of the light that falls on them.

If the observer stands on a lofty mountain peak overlooking a broad region of much lower level, the horizon reds are intensified. This is due to the more complete scattering away of the blue rays in the additional measure of dusty air traversed by the horizon beams, which to the observer on the lowlands have to pass through a less distance.

Refraction has a small share in producing sunset colors. When the rays of light enter the atmosphere obliquely, they are bent or refracted from their path towards the denser lower air. An observer always sees the sun at a greater altitude above the horizon than it really is. When the sun appears to be on the horizon line, it has really passed below the horizon. Refraction is greatest at the horizon, when it increases the apparent altitude of a celestial object by about half a degree. The sun's angular diameter being of the same measure, it follows that the lower limb or edge of the solar disc will seem to rest on the horizon when the upper limb is really just passing below it.

The longer-waved rays are refracted less than the finer-waved ones; a star seen in a telescope near the horizon exhibits the prismatic colors, with red below and blue above. After sunset every solar beam will be similarly broken into a short vertical spectrum. The successive spectra will overlap, and aid

the process of selective scattering in producing a gradation from red on the horizon through yellows to blue in the upper sky; but of the two processes, the scattering is the more effective.

It does not appear warrantable to attribute any considerable share of atmospheric colors to selective absorption and transmission. There is no sufficient direct evidence to show that either dry air or water vapor are colored, in the sense that chlorine is colored. The variation in the intensity of the sunset colors proves that they cannot be due to absorption by the air itself. If it be assumed that the red of sunset is caused by absorption of the blue rays by water vapor, then the blue of the open day-time sky remains to be explained. Moreover, while the intensity of sunset reds increases with the dampness of the lower atmosphere, it does not show a close dependence on the absolute amount of water vapor present; the colors vary with the approach of the vapor to the condition of saturation and condensation. It therefore seems more reasonable to disregard the absorptive effects of water vapor in the gaseous state, and consider only the action of minute particles of condensed water or ice in aiding other kinds of suspended matter to cause sunset colors by scattering the solar rays.

69. The twilight arch. The colors on the horizon opposite the sun are also best explained by selective scattering. At sunset the pink band along the eastern horizon, which rises and forms the twilight arch as the sun descends, is the return to us of the excessively red light by which the eastern sky is then illuminated; the light being red when it reaches us, it becomes redder still as it goes further on, and is then returned as a faint illumination from the particles that it encounters. It should be recalled in this connection that the backward scattering of light is symmetrical with the forward scattering; and that the red rays are returned backward in greater force than the blues, which are for the most part thrown off laterally. As the light from the western horizon is red as it passes us, it is still more reddened, although diminished in intensity, by the selective scattering that returns it from the eastern sky. At the same time many irregular scatterings give us blue light with the red, and thus make the arch of a rosy color, rather than an intense red as in the west. A similar rosy color is seen opposite sunset on the snow of mountains or on lofty clouds, as in the rear of a thunder-storm retreating in the east after sunset; the snow or cloud does not produce the red color, but simply sends back to us the color that falls on it.

The dark bluish area beneath the rising twilight arch is the shadow of the earth on the sky. The rays of light from the sun cannot reach this portion of the sky without many turnings and scatterings on the way, so that any light coming back to the eye from the shaded sky has lost nearly all its red, and hence appears of a distinct blue color. The under edge of the arch is

distinct at first, but becomes blurred as it rises. This is because our line of sight after sunset departs more and more from the surface which separates the arch and shadow. Just after sunset we stand almost in the surface of separation and look closely along it; then there is a sharp separation between the colors above and below it: but as the rotation of the earth carries us into the shadow, we look across the surface and our line of sight then receives rays from both the blue shadow and from the rosy area; hence the colors are blended and their separation fades away.

Subordinate effects of the same kind as the blue shadow of the earth are seen in shadows of clouds and mountains on the sky. When the western sky contains massive clouds at sunset, the eastern twilight arch will be distinctly interrupted by delicate bluish rays, whose narrow lower ends all converge to a point on the edge of the arch opposite to the sun; the convergence being an effect of perspective on really parallel cloud shadows. In the same way, if an observer stand upon a lofty mountain at sunset, he will see the shadow of the mountain rising above the eastern horizon and interrupting the twilight arch. The shadows of adjacent peaks are also sometimes seen, but less distinctly. The isolated summit of Pikes Peak in the Rocky Mountains, or of Fujiyama in Japan, casts an immense solitary conical blue shadow on the sky at sunrise or sunset.

70. Sunset and sunrise glows. The white or yellow oval glow, changing to rose or purple as it fades away after sunset, is accounted for by the interference of solar rays scattered or diffracted from particles of various sizes in the lower and upper air.

The bright white glow around the sun at noon is essentially a diffraction glow on particles of many sizes. Near sunset the solar rays pass through so great a thickness of air and encounter so many more particles than at noon that the disc becomes brighter and much larger; but the particles encountered are still of so many different sizes that no distinct color is produced.

Shortly after sunset, when the observer and the air for several thousand feet above him are in the shadow of the earth, the glow comes only from particles in the upper air; and as these are small and of more nearly uniform size than those near the earth, the glow increases still more in purity of color as the lower air darkens, and the delicate rosy marginal color makes its appearance at an angular distance of 20 or 25 degrees from the sun. The color is not brilliant, but in the waning twilight it is clearly seen; in fine weather it constitutes one of the chief glories of the western sunset sky. It descends and fades away when the sun is about six degrees below the horizon; to be followed in the clearest weather by a much fainter rosy or purple afterglow, visible for a short time. This is best explained as a second ring of the same origin as the first, but at a greater angular distance from the sun. Its

suggested explanation by reflection is not satisfactory, because there are no particles in the lofty layers of the atmosphere then illuminated that are large enough to cause reflection; they can only diffract the light that falls on them. A moderate haze in the lower air weakens or obscures the rosy glows. They are therefore in general better seen in winter than in summer.

71. The red sunsets of 1883-84. In 1883, '84, '85, the glows obtained an extraordinary development, which is believed to have resulted from the presence then in the upper air of minute particles of dust or of condensed vapor, blown out of the volcano, Krakatoa, in the Strait of Sunda, between Java and Sumatra, late in August, 1883. At the same time the sun was surrounded even at noon on clear days by a dusky reddish ring of about 20° radius, known as Bishop's ring, after its first observer; this being the daytime appearance of the diffractive sunset glow, then so brilliant as to be visible in the fully illuminated sky, but usually of faint intensity, so that it can be seen only after sunset.

These brilliant sunsets attracted great attention at the time, and the records of their appearance in different parts of the world have been carefully studied. The eruption occurred with excessive violence on August 26 and 27, 1883, destroying half of the island of Krakatoa, leaving water more than a thousand feet deep where the volcano had stood before, and shaking the air so vigorously as to produce an atmospheric wave that broke windows a hundred miles away and travelled around the earth, converging at the antipodal point and then returning to its source; the automatic barometric records kept in different parts of the world indicate that the wave went out from Krakatoa and back from the antipodal point at least three times. The sounds of the explosion were heard over the Malay archipelago, half of Australia and half of the Indian ocean, even three thousand miles away. The sea-waves driven away from the bursting volcano caused great destruction on the coasts of the neighboring islands, drowning over 30,000 persons, and then swept across the oceans, registering their arrival on tide-gauges in various harbors in all parts of the world. Pumice and dust blown from the volcano blackened the sky and fell for hundreds of miles around, obstructing the sea. The finer dust and the icy particles condensed from the ejected vapor, whose sudden expansion is believed to have caused the explosion of the volcano, reached great altitudes in the atmosphere, and there spread around the world. As the dust spread over the sky, the sunset colors became extraordinarily brilliant, the usually faint second glow restoring vivid colors to the fading sky and exciting remark from all observers. The dates of the first occurrence of these striking phenomena in different parts of the world have been carefully charted, and it is thus seen that they spread rapidly westward from Krakatoa around the equator, completing the circuit of the earth in fifteen days, and then gradually

spreading poleward. Before the end of the year they were visible in all parts of the world. Their duration extended through the greater part of 1884 and into 1885, and the reddish ring around the sun was seen even in 1886. An excessive fineness of the suspended particles is thus indicated.

Brilliant sunsets were recorded in 1783 and in 1831, following great volcanic eruptions in those years; many other less remarkable examples of the same relation of sunset colors to volcanic eruptions have been noted. In recording such phenomena, it is important to note the dates of first visibility of the vivid colors, the time of the appearance and change of the successive colors, and their altitude above the horizon.

While an excess of very fine particles in the upper air increases the intensity of the sunset colors, an excess of coarser dust in the lower air reduces their brightness, and may even conceal them entirely. The delicate rosy glow is the first to be extinguished in this way, and as the turbidity increases even the stronger horizon tints fail to appear. The direct rays from the solar disc are the last to disappear; only the most intense red rays then reach the observer, leaving the sky a dull dead gray all around: sometimes the sun itself disappears in the hazy or smoky sky before it sets, even though no clouds are present.

A peculiar instance of the dependence of sky colors upon the dustiness of air was observed by the author several years ago in Cambridge, Mass. In the afternoon a brief squall of dry wind blew a great quantity of dust into the air. The sunset was devoid of colors except in the sun itself, which disappeared on approaching the horizon as a circle of deep crimson color. The cloudless sky was dull; and even as darkness came on, few stars appeared. The half moon, well up in the sky, shone out with a very unusual red color; at the same time some of the brighter stars appeared faintly, and with a reddish tinge. In the course of a few hours the quantity of dust was so greatly diminished, either by settling down or drifting away, that the stars appeared in the usual number and the moon lost its red color, passing gradually through orange to its normal tint.

72. Mirage and looming. Further account may be given here of certain peculiar effects of reflection on atmospheric layers of different density, of which brief mention was made in Section 46. Rays of light may be totally reflected at the surface of contact of two layers of unlike density, if the angle of incidence is very large. Hence when a layer of very warm air lies close to the surface of a plain, the eye of an observer who stands above this layer receives from the further parts of it only the reflected light of the sky or of elevated objects in the distance, and no rays from the ground beneath. Objects thus reflected are inverted, as if from a horizontal mirror whose plane is below the observer; such an appearance being called a mirage. It is com-

mon in calm weather and in the hot hours of the day on level desert surfaces, and also over water surfaces when a light wind from the land carries out air of a temperature unlike that of the water. A slight change in the height of the observer may cause a considerable change in the mirage; and if no mirage is seen a few feet above the ground or water, it may often be discovered at a less height. Some slight mirage is nearly always visible when the eye is within an inch or two of an extended surface of quiet water.

Over the sea in the neighborhood of the coast, and particularly in the Arctic regions, it often happens that the surface of reflection is above the observer. The reflection is then called looming, and is characterized by an inverted image above the object. The image is often elongated vertically, producing an appearance of spires and pinnacles of fantastic form. Objects that are below the observer's horizon may be thus brought to view.

CHAPTER V.

THE MEASUREMENT AND DISTRIBUTION OF ATMOSPHERIC TEMPERATURES.

Thermometry.

73. Thermometers. The explanations in the third chapter of the processes by which the air is warmed and cooled prepare us to take up the study of thermometry, or the determination of the temperature of the air, with good understanding.

A thermometer, or heat-measure, is an instrument by which the temperature of a body may be compared with certain adopted standards of temperature. The standards are the freezing and boiling points of pure water, boiling to take place under one atmosphere of pressure. For convenience of measurement, the intermediate temperatures between these wide extremes are graded, or divided into degrees; but, unfortunately, the different countries of the world have not adopted degrees of the same value. The scale of the Fahrenheit thermometer, commonly employed in this country and in Great Britain and her colonies, begins to count its degrees at a temperature not naturally defined; and places freezing at 32° and boiling at 212°. One hundred and eighty degrees, therefore, correspond on this scale to the difference between the temperatures of freezing and boiling water. The centigrade thermometer places its zero point at freezing and 100° at boiling; 100 centigrade degrees, therefore, equal 180 Fahrenheit degrees. When temperatures below the zero point are observed, they should be recorded with a minus sign, thus: —6°, and read "six degrees below zero."

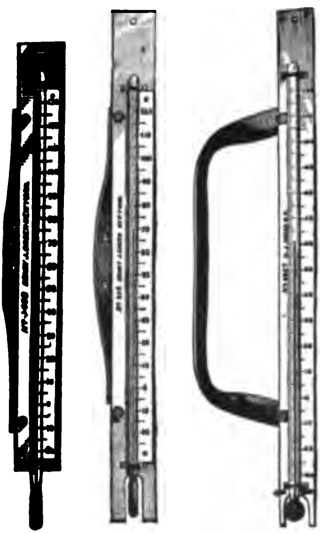

Fig. 7.

The essentials of a good thermometer may be summarized as follows: The liquid in its tube must not freeze at any temperature that it is likely to experience. The volume of the bulb must be large in comparison with that of the tube, in order to render any change of volume by expansion in the bulb easily apparent in the greater length of the column in the tube. The tube must be of constant diameter, in order that equal increments of temperature

shall produce equal increments of length in the column. The scale should be etched or cut on the tube itself. The temperatures indicated by the graduation of the scale should have been carefully determined by comparison with accurate standards, and the intermediate divisions of the scale must be accurately marked at equal intervals.

It is a waste of labor to undertake observations of temperature with an inaccurate thermometer. A great deal of labor has unfortunately been wasted in this way. Many of the records of temperature that have been kept for years with great patience are worse than useless, because, being inaccurate, they are misleading. A good thermometer costs but two or three dollars; if carefully used, it may last many years. But even a good thermometer will not give good records unless it is properly exposed. Persons who desire to undertake regular meteorological observations should therefore apply to their local State Weather Service, or to the Weather Bureau in Washington, for full instructions concerning instruments, shelters, records, etc.

It is desired that a thermometer, when read, shall indicate the temperature of the open air at a small height over the ground, and not so close to buildings as to be affected by them. For this reason, the thermometer should be placed in a shelter, with protection from sunshine and rain or snow, but well open to the wind. It should be removed, if possible, from buildings and trees. When this is not possible, a small shelter placed on the shady side of a building, not too high above the ground, may serve; but it is not worth while to attempt to make a record of temperature unless the exposure is such as will warrant confidence in the records. In cities, a shelter on the roof of a building is probably better than at a window in a narrow street. It is to be expected that the temperature of cities should be somewhat higher than that of the surrounding open country.

74. The sling thermometer. In case no satisfactory shelter can be provided, correct records can be obtained by tying the thermometer to a string two or three feet long and whirling it around in the air until its reading does not change. The instrument thus arranged is called the sling thermometer. It is of especial use in exploring expeditions, or in local studies of the variations of temperature in a small district, where no proper shelter can be counted on, and it should be frequently employed when establishing a permanent shelter for meteorological instruments, in order to determine the difference between its temperature and that of the surrounding air, particularly in quiet sunny weather.

Records of temperature should be carefully entered in a book kept for that purpose. At the beginning of the record a careful statement should be made describing the thermometer and its location; any subsequent change of exposure should be clearly entered at its proper date.

75. Thermographs. The temperature of the air is continually changing at a more or less rapid rate. A perfect record would present a complete indication of all the changes. This cannot be gained by ordinary thermometers, but it is practically gained by self-recording instruments, known as *thermographs*. Two styles of thermographs are illustrated in the accompanying figures.

The Draper self-recording thermometer, or thermograph, an American instrument (Fig. 8), possesses a metallic thermometer, one end of which is fixed while the other end is attached to a train of levers, to magnify the small movements due to expansion or contraction by change of temperature. The end of the last lever carries a pen which contains a non-freezing glycerine ink, and rests on a circular record sheet that rotates once a week. As the temperature rises, the pen is carried outwards from the center of the sheet; as the temperature falls, the pen is carried inwards. The sheet is divided into days and hours by curved radial lines, and into degrees by concentric circular lines; so that the temperature at any time can be easily read off. Although instruments of this kind are not so accurate as good mercurial thermometers, they make up for their slight inaccuracy by the continuity of their record; and if checked by frequent readings of a mercurial thermometer and driven by an accurate clock, they are of great value. The Draper self-recording thermometer is made by the Draper Manufacturing Co., 152 Front Street, New York. The cost of the instrument is $15.00.

FIG. 8.

The Richard Frères thermograph (Fig. 9), made in Paris (Glaenzer & Co., 80 Chambers Street, New York, are agents for this country), is of a somewhat different pattern, costing, without duty, about $25.00. The thermometer is here a flat bent tube of brass, containing a non-freezing liquid. One end of the tube is fixed to the frame of the instrument; the other end moves freely with change of temperature, and works a train of levers, which, as before, magnify the movement of the tube. If the temperature rises, the greater expansion of the liquid than of the tube bends the tube towards a straight line; if the temperature falls, the elasticity of the tube bends it into a sharper curve. The pen at the end of the last lever bears lightly on a sheet of paper that is wrapped around a drum or barrel; the drum is turned around once a week (or once a day, if so ordered) by clock-work within it. The pen rises and falls with the temperature and thus writes its record on the rotating drum.

Sample curves from sheets of this pattern are given on reduced scale in the adjoining figures, 10, 11, 12. Curve *a*, Fig. 10, illustrates a period of clear warming weather from April 27 to 30, 1889, with large and regular diurnal range, from a record kept by the Jackson Company at Nashua, N. H., for the

New England Meteorological Society; curve *b* presents the effects of a spell of cloudy weather at Nashua accompanying the passage of the great Hatteras hurricane of September 13 to 16, 1889; curve *c* shows the change from a time

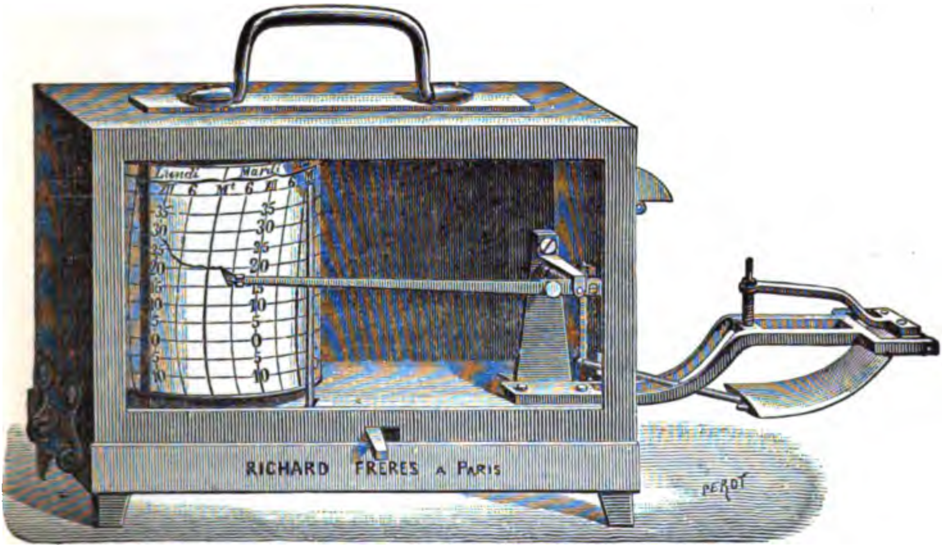

FIG. 9.

of moderate winter weather to a cold spell at Nashua, February 22 to 25, 1889, the change occurring at midnight of the 23–24; curve *d* exhibits a steady fall of temperature from the night of one day over the next noon to the following night, in the coming of a winter cold spell at Nashua, January 19 to 21, 1889; curve *e* is the reverse of the preceding case, being the effect of an

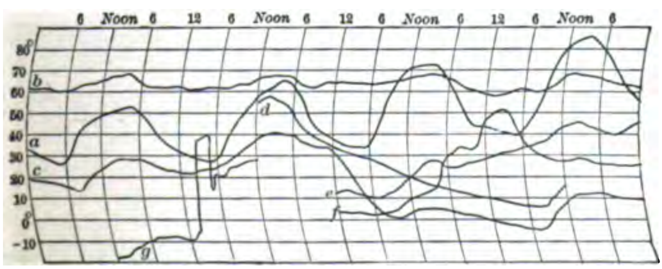

FIG. 10.

approaching mild spell at Nashua, December 16 and 17, 1888, in which there was a continuous rise of temperature through a night from noon to noon; curve *f* illustrates especially well the value of thermograph records, in giving

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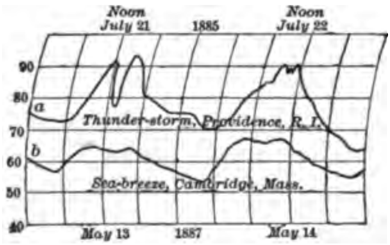

FIG. 11.

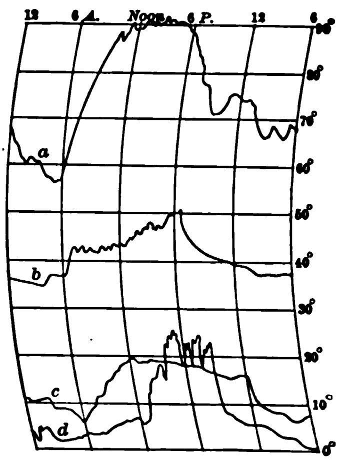

FIG. 12.

76. Maximum and minimum thermometers. When thermographs cannot be employed, other devices are introduced in order to secure special records in the simplest way. The most important of these are seen in the maximum and minimum thermometers, shown in Fig. 13. These instruments register the highest and lowest temperatures of the day. The maximum thermometer, the lower one in the figure, has a narrow constriction in the tube just outside of the bulb. As the temperature rises, the mercury is driven out of the bulb; but as the temperature falls, the mercury does not return; the upper end of the column in the tube therefore registers the highest reading since the instrument was set. Setting is done by whirling the instrument rapidly on a peg at its head, when the mercury is driven back past the constriction into the bulb. The minimum thermometer contains a transparent non-freezing liquid, instead of mercury. A small, double-headed pin or index lies inside of the liquid in the tube (between 60° and 70° in the figure). By raising the bulb, the index slides along to the end of the liquid column; the

instrument is then left, slanting gently towards the bulb; as the temperature rises, the expansion of the liquid is too slow to push the index along with it; but when the temperature falls below that at which the instrument was set, the surface cohesion of the liquid carries the index down the tube. The position in which the upper end of the index lies therefore registers the lowest or minimum reading since the previous setting. A pair of good maximum and minimum thermometers costs about $6.00; they are very easily

FIG. 18.

cared for; a single observation, as late in the evening as possible, sufficing to determine the diurnal range of temperature, which is one of the most important weather elements. The time at which the highest and lowest temperatures occurred are, however, not indicated. The readings of the maximum and minimum thermometers should always be entered in the record book before the instruments are set for a new observation.

77. **Black-bulb thermometer.** It is sometimes desired to obtain an indication of the intensity of sunshine, independent of the temperature of the air. This is roughly effected by exposing a maximum thermometer having the bulb coated with dull lamp-black, the thermometer being enclosed in a glass tube from which the air has been exhausted. The lamp-black on the bulb absorbs a large share of the sunshine, and the absence of air around the bulb prevents cooling by conduction. A temperature much above that of the surrounding air is thus reached. It is customary to record simply the excess of the maximum thus gained over that of the ordinary maximum reading. This excess, however, varies so greatly with the conditions surrounding the instrument that it is not admissible to regard observations with black-bulb thermometers as having any precise or comparable value. The instrument cannot be recommended for ordinary observations.

78. **Record of temperature: mean temperatures.** Besides the record of the temperature of the air at certain times, it is desired to determine also the mean temperature of the air for the day, the months and the year. This could be done by making hourly records and calculating their mean for each day; the average of the successive diurnal means would give the mean of the month; and the average of the monthly means, the mean of the year. The

average of twenty or thirty successive yearly means suffices to establish the mean temperature of a place.

It is manifest, however, that few persons can maintain a long series of hourly observations. These are sometimes taken at the larger observatories; but they are now mostly superseded by thermographs, checked by maximum and minimum thermometers. The question then arises, at what several hours of the day shall ordinary observations be taken in order that their average shall give a close measure of the true diurnal mean? This is determined by the hourly observations that have been made at certain stations. First, the mean temperatures of the successive hours, 1 o'clock, 2 o'clock, 3 o'clock, etc., are separately determined for every month. Selection is then made of certain pairs or groups of hours whose mean corresponds closely to, or differs by a small number of degrees from the true diurnal mean. While the mean of two, three or four observations in a day at the hours thus chosen cannot be trusted to determine the true mean of any single day, yet if the observations are continued for a month, they will serve to determine accurately enough the monthly mean, and then from successive monthly means the annual mean may be derived. Hours thus recommended for observation are as follows: — For two daily records, hours of the same name in the morning and afternoon, as 7 a. m. and 7 p. m.; 8 a. m. and 8 p. m., etc.; for three daily records, 6 a. m., 2 and 9 p. m.; or 7 a. m., 2 and 9 p. m. (add double the reading at 9 p. m. to the readings at the other hours and divide the sum by four). The mean of the maximum and minimum records is often used, but it is from a half to one degree too high.

None of these combinations have a greater error than a degree or a degree and a half in defining the monthly mean; and the error of the annual mean is still less. Tables published by the Weather Bureau and by the Smithsonian Institution at Washington give the corrections by which the means thus determined can be best reduced to the true means; and when thus corrected, the monthly and annual results may be trusted to a small fraction of a degree.

79. Climatic data. Besides the means already mentioned, it is customary to determine certain other data in defining the climate of a place. The most important are: — Monthly and annual means; mean diurnal range for each month; means of successive five-day periods or pentads through the year for single years, and for the same periods in groups of five years, or lustra; means of five-year periods, or lustra, beginning with years whose dates end with 1 or 6; absolute extremes of temperature for every month; mean of the monthly extremes for successive years; average difference between successive daily means.

When observations of good thermometers, well exposed and regularly read, extend over a lustrum period, they should be thus reduced, and the results

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In a region whose mean annual temperature is as [illegible] as in the central and eastern United States, it will be [illegible] that the values of successive [illegible] do not agree [illegible] [illegible] for the determination of the true mean annual temperature. It is therefore unadvisable to compare the mean temperatures of adjacent stations if the periods of observation do not agree. For example, the mean temperature of New Bedford, Mass., for the [illegible] beginning with [illegible] was [illegible]; that of Providence, R. I., for the [illegible] beginning [illegible] was [illegible]; from which it would appear that the mean of the latter was [illegible] higher than that of the former. But in the [illegible] beginning in [illegible] New Bedford had a mean of 49°.9; and in the [illegible] beginning in [illegible] Providence had a mean of [illegible]; and from this it would appear that the former was [illegible] above the latter. For the period [illegible] the two means differ less than half a degree.

Whenever possible, climatic comparisons of temperature or other data should be made for identical periods. If the observations for the region concerned do not cover the same period, it is desirable that they should be reduced to a definite period, such as an interval from [illegible] to [illegible]. This can be done approximately, as follows: — A certain station S may have records of temperature from 1855 to 1875; determine the mean for this period at adjacent stations of long continued observation; determine the average difference between these means and the means for the period 1870–1890; apply this average difference as a correction to the mean of station S; and the result will be the probable mean for its locality for the period 1870–1890.

80. Isothermal charts. Observations of temperature have been maintained at many stations in all parts of the world, some of them for all the years of this century, but generally for shorter periods; the distribution of temperature over the earth is studied by means of such records. In order to make observations taken at different altitudes on land comparable with one another, it is customary to reduce all temperatures to sea-level by adding one degree to the annual or monthly mean for every three hundred feet of altitude; but in preparing daily weather maps, the actual thermometer readings are charted.

Observations made at sea are gathered by the Hydrographic Offices of various countries; they are classified first by position, all of those taken within a certain "square" of latitude and longitude being placed together; and again by months. The records are then reduced to the true mean of the month as carefully as possible. As the variation of temperature at sea is generally much smaller and more regular than on land, the results thus obtained are fairly accurate, particularly in those parts of the ocean where

many vessels pass. The North Atlantic is especially well known in this respect. The "square" bounded by 20° and 25° longitude west of Greenwich and by 0° and 5° north latitude has 10,329 observations for March on the meteorological charts published by our Hydrographic Office at Washington.

When observations are gathered and reduced for many stations in various parts of the world, they may be charted on maps for the better illustration of the distribution of temperature; either for the mean annual temperature or for the mean of the seasons or of the months. Lines may then be drawn through places estimated to have mean temperatures of 50°, 60°, 70°, and so on; such lines are called *isotherms*, or lines of equal temperature. An isothermal chart thus constructed shows at a glance the areas that are on the average warmer or colder than any given temperature.

The best general isothermal charts of the world are those prepared by Dr. Julius Hann of Vienna, and by Professor Alexander Buchan of Edinburgh. The former are published in Berghaus' Physical Atlas [1] (1887); they present the isotherms on the centigrade scale. The latter include a beautiful series of monthly isothermal charts on the Fahrenheit scale, published in 1889 by the British government to illustrate an essay on the Atmospheric Circulation in the Report on the Challenger Expedition; but their high cost places them beyond general use. The isothermal charts on a small scale here presented are reduced from certain ones of this series; [2] the following sections call attention to the chief facts to be learned from them.

Distribution of Temperature over the Earth.

81. Contrast between the equator and poles. The most general facts presented by the isothermal chart for the year (Chart I) are the familiar high temperatures around the equator and the low temperatures about the poles. The sufficient reason for this has already been found in the greater annual value of insolation at the equator, decreasing to smaller values at the poles. The line of highest mean annual temperature, which may be called the mean annual heat equator, is not of uniform temperature all around its circuit. Its temperatures are five or more degrees higher on the lands than on the oceans. At the first glance one might explain this as the result of the lower specific heat of the land and of its non-volatile character: but as the inequality appears in the mean annual temperature, this explanation will not hold. It is true that if the mean temperature of the daytime or of the summer only were charted, the air over the lands would then be found on the average of higher temperature than that over the ocean for the above reasons; but as the mean for the year includes the conditions for night as well as for day, and for winter

[1] The meteorological section of this atlas, containing 12 charts, may be bought separately.

[2] These charts are on Gall's projection, in which the distortion of high latitudes is less than in Mercator's projection, commonly employed.

as well as for summer, the rapid cooling of the land and of the air close to it at the colder times must counterbalance the rapid heating in the warmer times; and hence for the mean of the year there should not be, for the suggested reasons, any higher temperature on the heat equator over the lands than over the oceans.

The true cause of the varying temperature along the heat equator is to be found in the interchange of polar and equatorial waters by the ocean currents, whereby the equatorial ocean is somewhat cooled and the polar oceans are much warmed; while on the lands there is no such interchanging process. The torrid lands are therefore hotter than the ocean of the same latitude; and the lands of high latitudes are colder than seas alongside of them. The lands take a temperature proper to their latitude, while the oceans attempt to equalize the temperatures between equator and poles.

A marked consequence of this is seen in the more rapid decrease of temperature from the mean annual heat equator towards the poles on land than on water; in other words, the poleward temperature gradient is stronger on the continents than on the oceans. Beginning, for example, in southern India and tracing a line almost northward to the Arctic coast of northeastern Siberia, the temperature falls from 85° to 0°, a decrease of almost a degree and a half of temperature in a degree of latitude. Following a northward line of the same length in the Atlantic ocean, the decrease is from 83° to 25°, or at a rate of a degree of temperature to a degree of latitude.

82. Irregularity of annual isotherms. The explanation that has already been given of the distribution of insolation over the earth might lead us to expect that the mean annual isotherms should coincide with the lines of latitude. A glance at the map shows that in many parts of the world the isotherms are unsymmetrical in the two hemispheres and that they depart greatly from an east and west course. We will first consider the character of the departures and then look for their explanation.

The unsymmetrical arrangement of the isotherms on either side of the geographical equator is first seen in the location of the heat equator in the northern hemisphere for the greatest part of its circuit, as may be shown by drawing a line bisecting the space between the pairs of corresponding isotherms of the torrid zone in either hemisphere. This line falls into southern latitudes only in the western part of the Pacific and in Australasia; its location here being due to a southern movement of the warm equatorial waters in that region, and to the higher mean temperatures of the land areas of Australasia than of the water areas on the opposite side of the equator. Elsewhere, the heat equator lies in northern latitudes. The absence of any southern continent to balance the effect of Asia explains its course across the northern Indian ocean; and, as will be seen in a later section, the inflow of

cool waters from the far southern latitudes displaces the line to the north of the equator in the eastern Atlantic and Pacific.

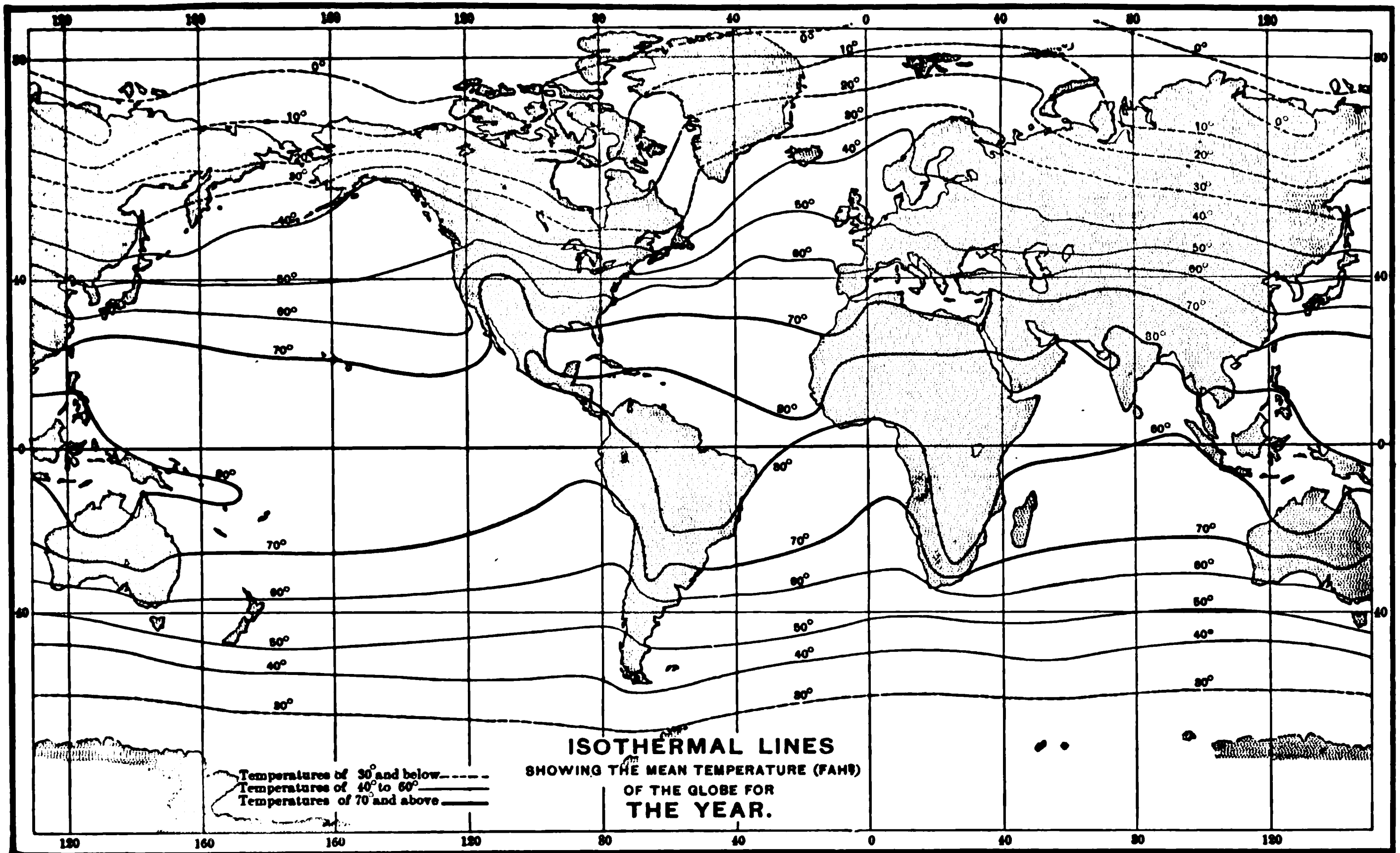
In the next place, the departures are small in the southern hemisphere, where the isotherms are remarkably regular, especially on the broad oceanic areas. The lines are much more irregular in the northern hemisphere; but in both hemispheres there are certain systematic deflections of isotherms from the parallels of latitude in passing from an ocean over a continent to the next ocean. In crossing eastward over South America, for example, the lines turn equatorward on the Pacific near the western coast; they loop strongly poleward in crossing the land, and finally run slowly towards the equator again in traversing the Atlantic. A similar irregularity, but not so pronounced, is seen in the passage of the lines over Africa and Australia. Coming now to our hemisphere, the isotherms on the Pacific turn somewhat towards the equator in the middle and lower latitudes as they approach North America; then entering the continent, they loop poleward; and on reaching the Atlantic they turn slowly toward the equator on their way across to Africa. A similar irregularity may be perceived, but less distinctly, on the broad lands of the old world in equivalent latitudes.

In the higher latitudes of the northern hemisphere the isotherms show just the opposite deflections. They turn towards the pole in the northeast Pacific, towards the equator in northeast America, strongly towards the pole in the northeast Atlantic, and so on. There is no land far enough south in the other hemisphere to exhibit deflections of this kind, except a small part of South America.

As a result of all this, the isotherms are crowded together on the eastern coasts of the northern continents, and spread far apart on the eastern side of the northern oceans. This is particularly apparent on the two sides of the Atlantic, where it is of especial interest, because the lands on either side of this ocean are at present the seat of the highest civilization in the world. In western Europe, one may travel a thousand miles northward without finding so great a change of mean annual temperature as would be found in a voyage of half that distance along our eastern coast.

The reason for the systematic deflections of the isotherms is found almost entirely in the even more systematic course of the great ocean currents. The interchange of ocean waters between the equator and poles already mentioned is performed in part by a surface flow towards the poles and a slow creeping of the deep waters back again towards the equator; but there is besides this an eddy-like circulation of the surface waters in the several oceans which is of much more importance in meteorology; because the temperature of the air, which we are now discussing, depends so largely on that of the surface on which it rests. The eddy-like currents of the ocean may now be simply generalized.

Chart I.



From Buchan's "Challenger" Report.

Bradley & Poates Engrs, N.Y.

75. Thermographs. The temperature of the air is continually changing at a more or less rapid rate. A perfect record would present a complete indication of all the changes. This cannot be gained by ordinary thermometers, but it is practically gained by self-recording instruments, known as *thermographs*. Two styles of thermographs are illustrated in the accompanying figures.

The Draper self-recording thermometer, or thermograph, an American instrument (Fig. 8), possesses a metallic thermometer, one end of which is fixed while the other end is attached to a train of levers, to magnify the small movements due to expansion or contraction by change of temperature. The end of the last lever carries a pen which contains a non-freezing glycerine ink, and rests on a circular record sheet that rotates once a week. As the temperature rises, the pen is carried outwards from the center of the sheet; as the temperature falls, the pen is carried inwards. The sheet is divided into days and hours by curved radial lines, and into degrees by concentric circular lines; so that the temperature at any time can be easily read off. Although instruments of this kind are not so accurate as good mercurial thermometers, they make up for their slight inaccuracy by the continuity of their record; and if checked by frequent readings of a mercurial thermometer and driven by an accurate clock, they are of great value. The Draper self-recording thermometer is made by the Draper Manufacturing Co., 152 Front Street, New York. The cost of the instrument is $15.00.

FIG. 8.

The Richard Frères thermograph (Fig. 9), made in Paris (Glaenzer & Co., 80 Chambers Street, New York, are agents for this country), is of a somewhat different pattern, costing, without duty, about $25.00. The thermometer is here a flat bent tube of brass, containing a non-freezing liquid. One end of the tube is fixed to the frame of the instrument; the other end moves freely with change of temperature, and works a train of levers, which, as before, magnify the movement of the tube. If the temperature rises, the greater expansion of the liquid than of the tube bends the tube towards a straight line; if the temperature falls, the elasticity of the tube bends it into a sharper curve. The pen at the end of the last lever bears lightly on a sheet of paper that is wrapped around a drum or barrel; the drum is turned around once a week (or once a day, if so ordered) by clock-work within it. The pen rises and falls with the temperature and thus writes its record on the rotating drum.

Sample curves from sheets of this pattern are given on reduced scale in the adjoining figures, 10, 11, 12. Curve *a*, Fig. 10, illustrates a period of clear warming weather from April 27 to 30, 1889, with large and regular diurnal range, from a record kept by the Jackson Company at Nashua, N. H., for the

New England Meteorological Society; curve *b* presents the effects of a spell of cloudy weather at Nashua accompanying the passage of the great Hatteras hurricane of September 13 to 16, 1889; curve *c* shows the change from a time

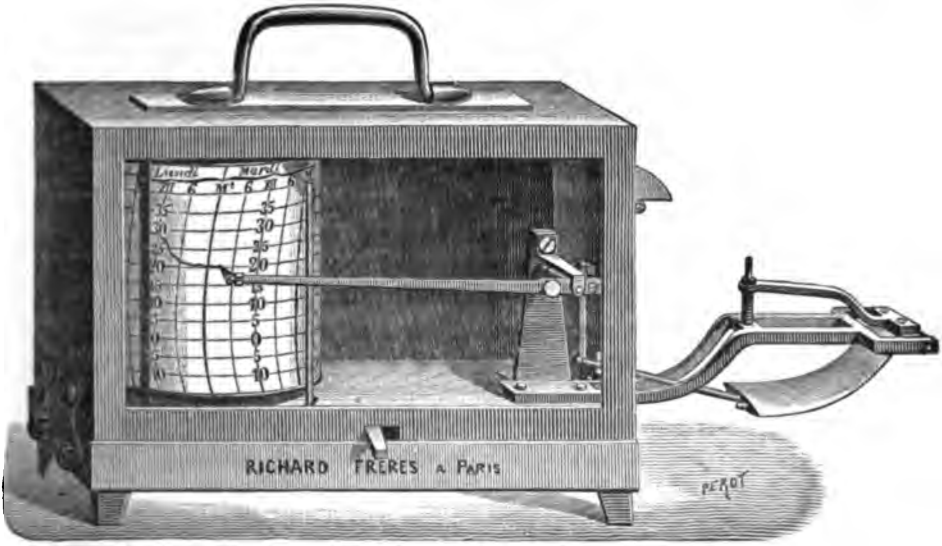
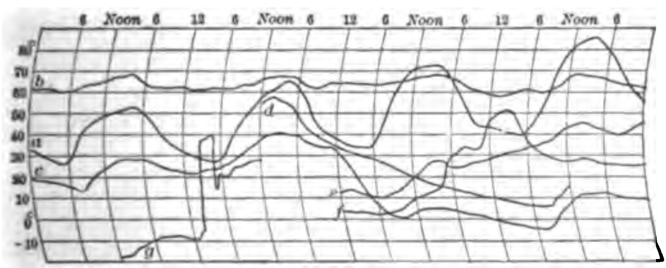


FIG. 9.

of moderate winter weather to a cold spell at Nashua, February 22 to 25, 1889, the change occurring at midnight of the 23–24; curve *d* exhibits a steady fall of temperature from the night of one day over the next noon to the following night, in the coming of a winter cold spell at Nashua, January 19 to 21, 1889; curve *e* is the reverse of the preceding case, being the effect of an



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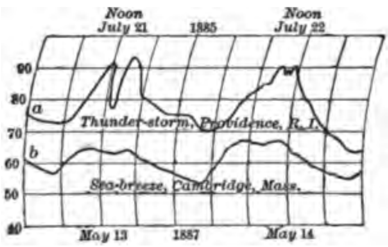

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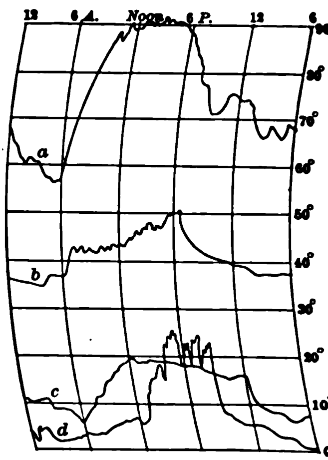

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The eddy of the South Indian ocean is similar to that of the South Pacific in being confluent with the great [illegible] on its polar side. Contrary to the representation [illegible] it does not give out a [illegible] around the southern end of Africa. The currents of the Northern Indian ocean are anomalous in changing their course with the seasons. In the northern summer they [illegible] normal left-to-right eddy, whose equatorial portion is then confluent with the corresponding portion of the South Indian eddy; in the northern [illegible] the eddy turns the other way, so that its equatorial portion then [illegible] to an equatorial counter-current.

The South Atlantic eddy is also confluent with the Antarctic [illegible] polar side, but it is strongly unlike the eddies of the other southern [illegible] giving out a great branch that flows obliquely across the [illegible] northern hemisphere. This is the result of the unsymmetrical [illegible] and South America, the former extending to the west, north [illegible] the latter extending to the east, south of the equator [illegible] sphere loses and the northern hemisphere gains a [illegible] water by this peculiar arrangement of continents and [illegible] be made of a cold current that wedges along the [illegible]

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Observations made at sea are gathered by the Hydrographic Offices of various countries; they are classified first by position, all of those taken within a certain "square" of latitude and longitude being placed together; and again by months. The records are then reduced to the true mean of the month as carefully as possible. As the variation of temperature at sea is generally much smaller and more regular than on land, the results thus obtained are fairly accurate, particularly in those parts of the ocean where

many vessels pass. The North Atlantic is especially well known in this respect. The "square" bounded by 20° and 25° longitude west of Greenwich and by 0° and 5° north latitude has 10,329 observations for March on the meteorological charts published by our Hydrographic Office at Washington.

When observations are gathered and reduced for many stations in various parts of the world, they may be charted on maps for the better illustration of the distribution of temperature; either for the mean annual temperature or for the mean of the seasons or of the months. Lines may then be drawn through places estimated to have mean temperatures of 50°, 60°, 70°, and so on; such lines are called *isotherms*, or lines of equal temperature. An isothermal chart thus constructed shows at a glance the areas that are on the average warmer or colder than any given temperature.

The best general isothermal charts of the world are those prepared by Dr. Julius Hann of Vienna, and by Professor Alexander Buchan of Edinburgh. The former are published in Berghaus' Physical Atlas [1] (1887); they present the isotherms on the centigrade scale. The latter include a beautiful series of monthly isothermal charts on the Fahrenheit scale, published in 1889 by the British government to illustrate an essay on the Atmospheric Circulation in the Report on the Challenger Expedition; but their high cost places them beyond general use. The isothermal charts on a small scale here presented are reduced from certain ones of this series; [2] the following sections call attention to the chief facts to be learned from them.

Distribution of Temperature over the Earth.

81. Contrast between the equator and poles. The most general facts presented by the isothermal chart for the year (Chart I) are the familiar high temperatures around the equator and the low temperatures about the poles. The sufficient reason for this has already been found in the greater annual value of insolation at the equator, decreasing to smaller values at the poles. The line of highest mean annual temperature, which may be called the mean annual heat equator, is not of uniform temperature all around its circuit. Its temperatures are five or more degrees higher on the lands than on the oceans. At the first glance one might explain this as the result of the lower specific heat of the land and of its non-volatile character: but as the inequality appears in the mean annual temperature, this explanation will not hold. It is true that if the mean temperature of the daytime or of the summer only were charted, the air over the lands would then be found on the average of higher temperature than that over the ocean for the above reasons; but as the mean for the year includes the conditions for night as well as for day, and for winter

[1] The meteorological section of this atlas, containing 12 charts, may be bought separately.

[2] These charts are on Gall's projection, in which the distortion of high latitudes is less than in Mercator's projection, commonly employed.

as well as for summer, the rapid cooling of the land and of the air close to it at the colder times must counterbalance the rapid heating in the warmer times; and hence for the mean of the year there should not be, for the suggested reasons, any higher temperature on the heat equator over the lands than over the oceans.

The true cause of the varying temperature along the heat equator is to be found in the interchange of polar and equatorial waters by the ocean currents, whereby the equatorial ocean is somewhat cooled and the polar oceans are much warmed; while on the lands there is no such interchanging process. The torrid lands are therefore hotter than the ocean of the same latitude; and the lands of high latitudes are colder than seas alongside of them. The lands take a temperature proper to their latitude, while the oceans attempt to equalize the temperatures between equator and poles.

A marked consequence of this is seen in the more rapid decrease of temperature from the mean annual heat equator towards the poles on land than on water; in other words, the poleward temperature gradient is stronger on the continents than on the oceans. Beginning, for example, in southern India and tracing a line almost northward to the Arctic coast of northeastern Siberia, the temperature falls from 85° to 0°, a decrease of almost a degree and a half of temperature in a degree of latitude. Following a northward line of the same length in the Atlantic ocean, the decrease is from 83° to 25°, or at a rate of a degree of temperature to a degree of latitude.

82. Irregularity of annual isotherms. The explanation that has already been given of the distribution of insolation over the earth might lead us to expect that the mean annual isotherms should coincide with the lines of latitude. A glance at the map shows that in many parts of the world the isotherms are unsymmetrical in the two hemispheres and that they depart greatly from an east and west course. We will first consider the character of the departures and then look for their explanation.

The unsymmetrical arrangement of the isotherms on either side of the geographical equator is first seen in the location of the heat equator in the northern hemisphere for the greatest part of its circuit, as may be shown by drawing a line bisecting the space between the pairs of corresponding isotherms of the torrid zone in either hemisphere. This line falls into southern latitudes only in the western part of the Pacific and in Australasia; its location here being due to a southern movement of the warm equatorial waters in that region, and to the higher mean temperatures of the land areas of Australasia than of the water areas on the opposite side of the equator. Elsewhere, the heat equator lies in northern latitudes. The absence of any southern continent to balance the effect of Asia explains its course across the northern Indian ocean; and, as will be seen in a later section, the inflow of

Although the sunshine is truly stronger during the southern summer on account of our nearness then to the sun, so great a share of the southern hemisphere possesses a water surface that the temperature, even under stronger sunshine, cannot rise to a high degree: in the northern summer, in spite of the slightly weaker sunshine consequent upon our greater distance from the sun, the temperatures produced are high, because it is so easy to raise the temperature of the land surface, and upon this the temperature of the air depends.

It may be also noted that if we take the mean temperature of the earth as a whole, it is not constant throughout the year. It might at first be expected that a slightly higher mean temperature should prevail in perihelion than in aphelion; and this would be true if the surface of the earth were of but one kind, or if the land and water were symmetrically distributed on either side of the equator; but under existing conditions it is found that the average

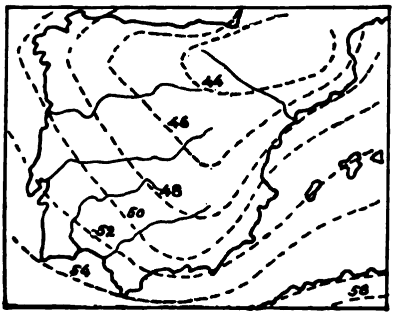


FIG. 14 (January).

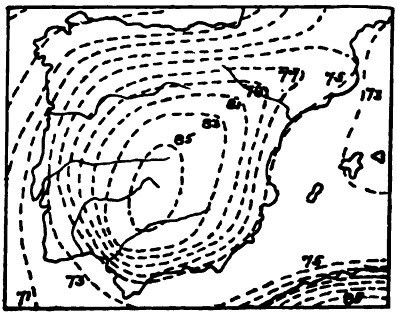
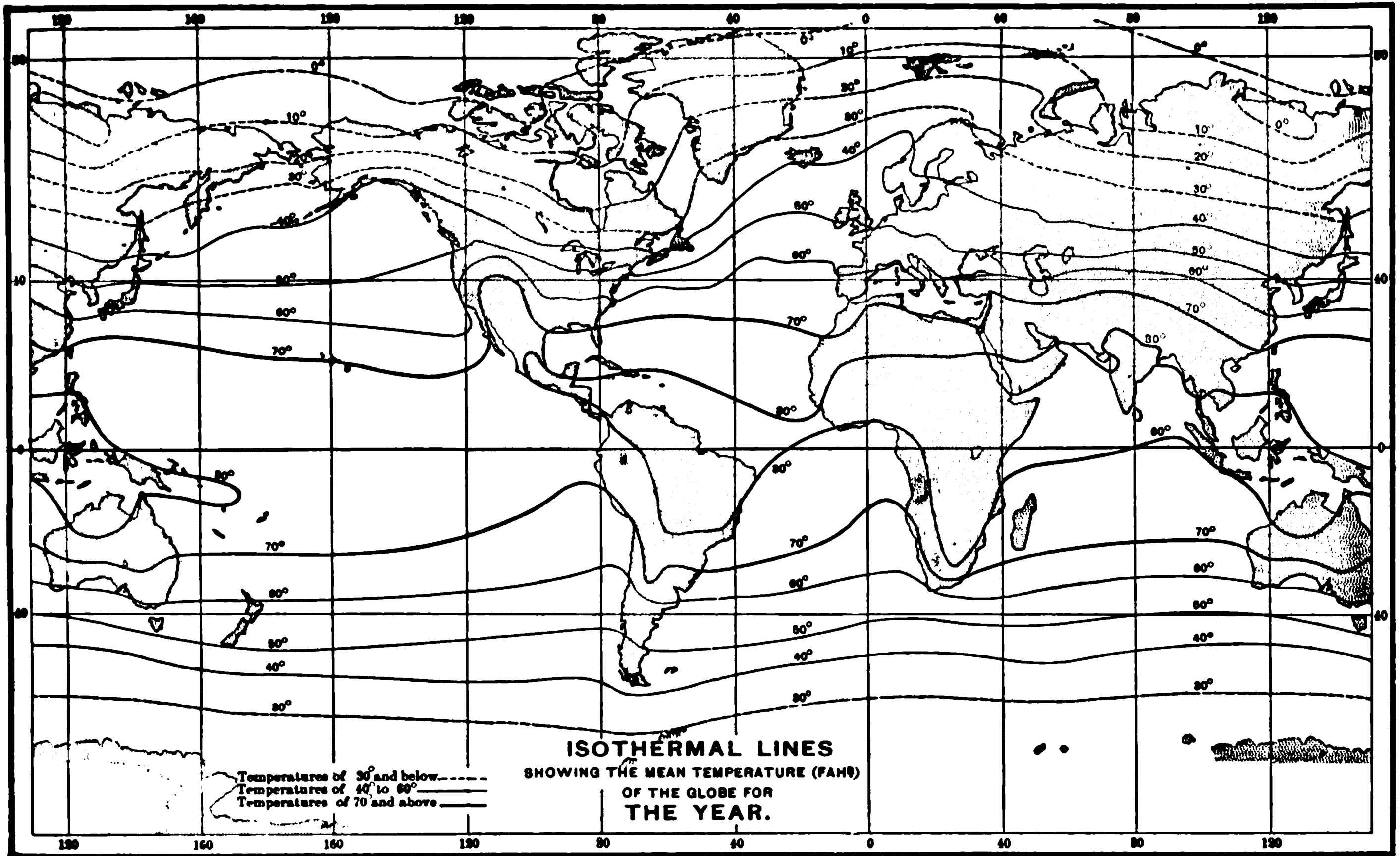


FIG. 15 (July).

temperature of the earth as a whole is higher in July, 63°, when the northern hemisphere is excessively hot and the southern hemisphere but moderately cold, than in January, 55°, when the southern hemisphere is moderately warm and the northern hemisphere is excessively cold.

Just as the summer of the northern or land hemisphere is warmer and the winter is colder than the same seasons of the southern or water hemisphere, so the various continental areas are warmer in summer and colder in winter than the adjacent oceans of similar latitude. This is strikingly apparent in the case of Asia, where the excessive heat of Arabia, Persia and northern India in summer and the excessive cold of Siberia in winter offer the extreme examples of terrestrial temperature variation. Similar variations but of a more moderate range are found in Australia. Even the interior of Great Britain is warmer than the surrounding sea in summer and colder in winter; and peninsular Spain and Portugal exhibit winter and summer isotherms roughly following the coast line, as in Figs. 14 and 15.

Chart I.



From Buchan's "Challenger" Report.

Bradley & Poates Engrs, N.Y.

83. General scheme of ocean currents. The North Pacific is the simplest of all the oceans from having practically no connection with the adjacent polar waters. Its great eddy turns from left to right, consisting of an equatorial portion moving from east to west; of a western portion, known as the Japanese current, which passes the Japanese Islands northeastward; a northern portion traversing the ocean towards Alaska; and an eastern portion flowing along our western coast to the southeast, completing the circuit. There is a noticeable subordinate eddy turning around from right to left in the Bay of Alaska, and a small, cold southward current from Kamchatka towards Japan. No significant supply of cold water comes from the Arctic ocean through Bering strait to the Pacific.

The South Pacific ocean has a similar general eddy of rather greater size; but its circulation is from right to left; its western portion flows among the Polynesian islands and near New Zealand and Australia; it gives off a small branch north of Australia to the Indian ocean; its southern portion is confluent with the great Antarctic eddy which runs from west to east around the south pole. The member of the Pacific eddy that flows equatorward along the western coast of South America is known as the Peruvian or Humboldt current; it furnishes a greater share of cool water to the equator than is brought by any other current. The two equatorial portions of the North and South Pacific eddies are not perfectly confluent, but are separated by a somewhat irregular counter-current, running from west to east a little north of the equator, and carrying a body of warm water to the coast of Central America.

The eddy of the South Indian ocean is similar to that of the South Pacific in being confluent with the great Antarctic eddy on its polar side. Contrary to the representation generally given on maps of ocean currents, it does not give out a branch to the South Atlantic around the southern end of Africa. The currents of the Northern Indian ocean are anomalous in changing their course with the seasons: in the northern summer they possess a normal left-to-right eddy, whose equatorial portion is then confluent with the corresponding portion of the South Indian eddy; in the southern summer the eddy turns the other way, so that its equatorial portion then corresponds to an equatorial counter-current.

The South Atlantic eddy is also confluent with the Antarctic eddy on its polar side, but it is strongly unlike the eddies of the other southern oceans in giving out a great branch that flows obliquely across the equator into the northern hemisphere. This is the result of the unsymmetrical form of Africa and South America, the former extending to the west, north of the equator; the latter extending to the east, south of the equator. The southern hemisphere loses and the northern hemisphere gains a large volume of warmed water by this peculiar arrangement of continents and ocean. Mention should be made of a cold current that wedges along the eastern side of Patagonia

towards the equator, the only distinct cold current on the eastern side of a southern continent.

The North Atlantic possesses the most peculiar system of currents of all the oceans. Its normal eddy receives on the southern side the great branch given off by the South Atlantic eddy, and in turn it gives off from its northern side a great volume of water which passes northward beyond Norway, circles around the great gulf commonly called the Arctic ocean, and returns greatly chilled to supply the cold Labrador current which creeps down our eastern coast as far as Cape Hatteras. A small counter current of variable extent flows eastward between the North and South Atlantic eddies into the Gulf of Guinea; its greatest extension coming in the late northern summer. This is closely analogous to the counter current of the equatorial Pacific.

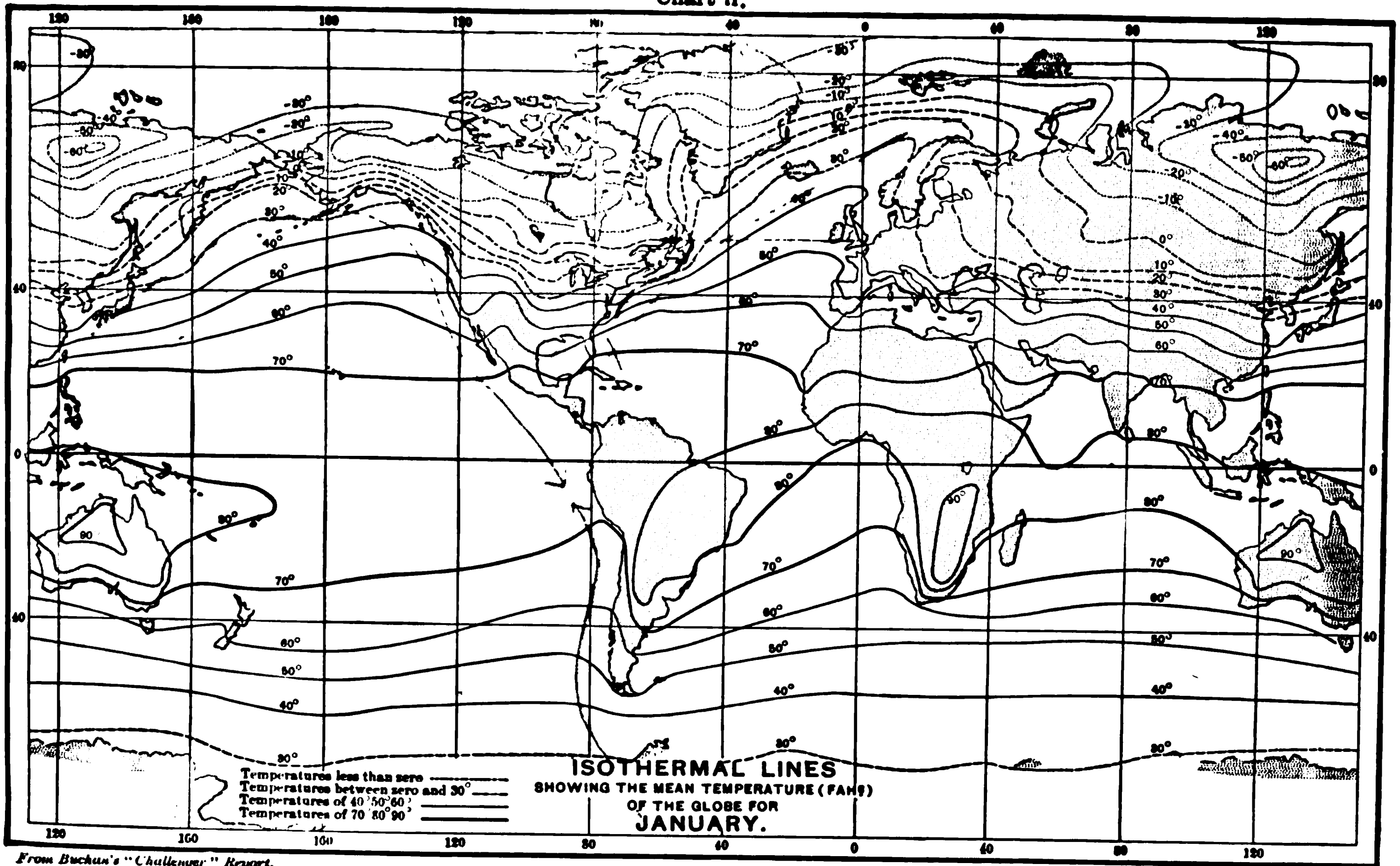
In the ordinary nomenclature of the currents of the North Atlantic, undue emphasis is given to that portion which is supposed to come from the Gulf of Mexico. It is true that a considerable volume of warm water issues from the Gulf between Florida and Cuba; the name "Gulf Stream" should be limited to this concentrated current. But it cannot be believed that all the warm water which flows northward off our eastern coast, and then drifts eastward and northeastward towards Europe, has issued from this moderate source. A considerable portion of it must have passed outside of the West India islands, without making the side circuit of the Caribbean sea and the Gulf of Mexico.

It should be noticed particularly that while the deformity of the North Pacific eddy on the north consists only in the subordinate eddies by Alaska and north of Japan, the North Atlantic eddy gives off a great branch on the north which makes the circuit of the Polar sea.

84. Deflection of isotherms by ocean currents. Bearing in mind that the equatorial waters are warm and the polar waters are cold, it follows that currents from the equator will tend to warm the air and deflect the isotherms towards the poles, while the currents returning from higher latitudes will carry the isotherms equatorwards. The deflections of the isotherms already described can all be accounted for by this principle.

The spreading apart of the isotherms on the western side of North America, for example, is due to the opposite courses of the return current that passes along California and Mexico in lower latitudes, and the subordinate eddy that circles from right to left around the Alaskan Bay in higher latitudes. Similarly opposite courses of ocean currents on a much larger scale are found in the eastern North Atlantic, and hence the divergence of the isotherms on the western coasts of Europe and northern Africa is still more marked; the equatorward deflections in low latitudes being controlled by the eastern portion of the normal North Atlantic eddy past Spain and the western Sahara, while the excessive northward deflections in the higher latitudes are

Chart II.



From Buchan's "Challenger" Report.

Bradley & Poates Engrs, N.Y.

due to the great Arctic branch of the North Atlantic eddy, so peculiar to this ocean and commonly considered an extension of the Gulf Stream.

The crowding of the isotherms on the eastern side of Asia depends in the lower latitudes on the northward turn of the Japanese current, and in the higher latitudes on the southward turn of the subordinate current between Kamchatka and Japan. The crowding of the lines east of North America is more pronounced, because the currents which control the isotherms there are of so remarkable a strength. Off the coasts of Florida and Carolina the isotherms are held to the north by the Gulf Stream proper; along the coast of New England and the Provinces they are carried strongly to the south by the powerful Labrador current.

In the southern hemisphere none of the continents offer serious interruption to the eastward course of the great Antarctic eddy; hence no strong deflections of the isotherms are found in high southern latitudes. Nearer the equator the deflections are similar to those of low latitudes in the northern hemisphere. The peculiar deflections of the isotherms in the northern hemisphere and their comparative regularity in the southern are thus well accounted for.

An interesting corollary may be drawn from these explanations. If our earth possessed a surface of level land only, the difference of temperature between the equator and poles would be much greater than it is at present, even greater than on the existing lands where the poleward decrease of temperature is comparatively rapid; while on a world of continuous water surface, if there existed a gradual interchange between equatorial and polar waters, such as might be reasonably expected, the contrast of equatorial and polar temperatures would be smaller. In either case the distribution of temperature all over the world would be as regular as we now find it in the southern hemisphere.

85. Isotherms for January and July. We may next examine Charts II and III, representing the mean temperatures of the earth in January and in July. The general variation from winter to summer in either hemisphere is simply enough explained by referring to the variations in the values of insolation with the seasons, as given in Section 27. It should be recalled in this connection that in January the earth is nearest to the sun; hence if distance from the sun alone controlled our temperatures, we should expect to find the southern summer, which occurs in perihelion, of higher temperature as a whole than the northern summer, which occurs in aphelion. Comparing the summers of the two hemispheres, we find that the reverse is true. The summer of the southern hemisphere in January is marked by moderately high temperatures; the summer of the northern hemisphere in July possesses excessive heats over large areas. The location of the latter areas being all on the land, we readily discover the reason for this perhaps unexpected result.

Although the sunshine is truly stronger during the southern summer on account of our nearness then to the sun, so great a share of the southern hemisphere possesses a water surface that the temperature, even under stronger sunshine, cannot rise to a high degree: in the northern summer, in spite of the slightly weaker sunshine consequent upon our greater distance from the sun, the temperatures produced are high, because it is so easy to raise the temperature of the land surface, and upon this the temperature of the air depends.

It may be also noted that if we take the mean temperature of the earth as a whole, it is not constant throughout the year. It might at first be expected that a slightly higher mean temperature should prevail in perihelion than in aphelion; and this would be true if the surface of the earth were of but one kind, or if the land and water were symmetrically distributed on either side of the equator; but under existing conditions it is found that the average

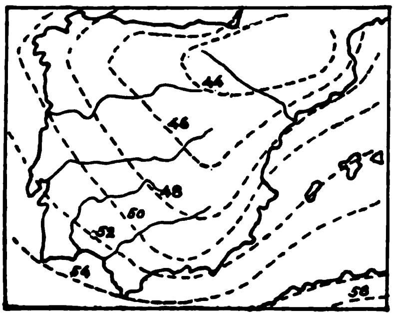


Fig. 14 (January).

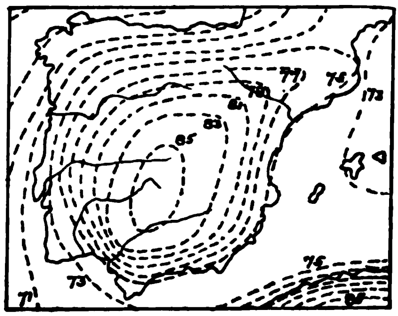
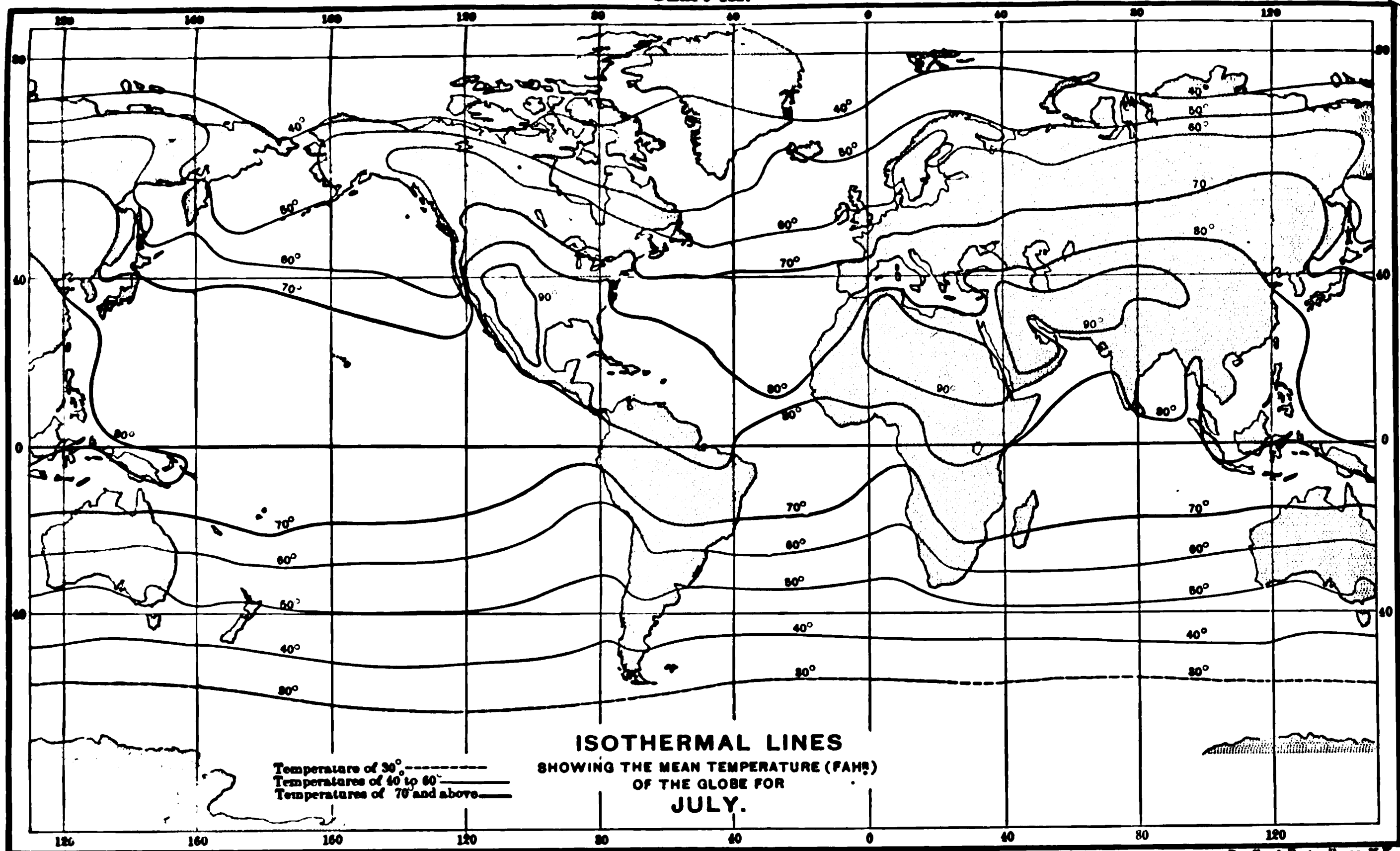


Fig. 15 (July).

temperature of the earth as a whole is higher in July, 63°, when the northern hemisphere is excessively hot and the southern hemisphere but moderately cold, than in January, 55°, when the southern hemisphere is moderately warm and the northern hemisphere is excessively cold.

Just as the summer of the northern or land hemisphere is warmer and the winter is colder than the same seasons of the southern or water hemisphere, so the various continental areas are warmer in summer and colder in winter than the adjacent oceans of similar latitude. This is strikingly apparent in the case of Asia, where the excessive heat of Arabia, Persia and northern India in summer and the excessive cold of Siberia in winter offer the extreme examples of terrestrial temperature variation. Similar variations but of a more moderate range are found in Australia. Even the interior of Great Britain is warmer than the surrounding sea in summer and colder in winter; and peninsular Spain and Portugal exhibit winter and summer isotherms roughly following the coast line, as in Figs. 14 and 15.

Chart III.



From Buchan's "Challenger" Report.

86. Poleward temperature gradients in winter. Looking again at the general rate of decrease of temperature from equator to pole, it will be seen that this is stronger in the winter hemisphere than in the summer hemisphere; particularly if the comparison is made between the northern winter and the southern summer. The change in the value of this temperature gradient is not well marked in the southern hemisphere, because there the change of temperature with the seasons is comparatively small; but in the northern hemisphere it is very distinct. In the summer time the great area of lands in high northern latitudes determines the occurrence of comparatively high temperatures in the far north; hence the poleward decrease of temperature at this time is comparatively gradual; but in the winter time the great area of northern lands allows the temperature to fall excessively low, and the decrease of temperature from the equator towards the north pole is extremely rapid. This will be found to be of importance in explaining the more violent winds of our winter season.

87. Migration of isotherms. The migration of the sun north and south of the equator, by reason of the obliquity of the earth's axis to the plane of its orbit, causes a migration of the heat equator also; but while the sun shifts its position from $23\frac{1}{2}°$ north to $23\frac{1}{2}°$ south of the equator, the heat equator generally migrates by a much less amount. Moreover, while the sun stands farthest north or south of the equator in June or December, the greatest migration of the heat equator northward or southward is found in July or August and January or February; just as the hottest part of the day is an hour or two after noon. The shifting of the heat equator is particularly small on the oceans. On the Pacific it moves over 15° or 20° of latitude; on the Atlantic still less, and only on the western part of this ocean does it cross to the southern hemisphere when the sun is south; being held elsewhere north of the equator by the cool African current: a notable consequence of this will be found in Section 225. On the continents it shifts over a greater distance. In Africa it moves from about 23° N. to 20° S.; this migration being almost symmetrical north and south, because the land there extends about equally on either side of the equator. In America the migration is from 35° N. to 15° S. A more peculiar case is found in the Indian ocean and on the land to the north of it. When the sun comes north of the equator, the heat equator runs beyond it to the deserts of Persia in latitude 33° N. In the opposite season, when the sun goes south, the heat equator hangs behind it and reaches only 10° S. latitude, because there are no southern lands in these longitudes to tempt it further; in Australia, however, it reaches latitude 20° S. Important consequences of this unsymmetrical migration will be found in the chapters on the winds and rainfall.

c13.5°

An interesting study may be made of the migration of any intermediate isotherm of the temperate zones. On the oceans of the southern hemisphere, the isotherm of 50° shifts annually from latitude 35° or 40° to 45° or 50°. In the middle of the North Pacific the shift of the 50° isotherm is from latitude 43° to 53°. On the axis of North America the same line migrates from Arkansas, latitude 33°, almost to the Arctic shore of British America, in latitude 67°; and over Asia it travels from latitude 28° in southeastern China to latitude 70° near the Lena delta in Siberia.

88. Northern winter isotherms. The irregularities in the course of the isotherms already examined on the chart for the year are exceeded by those found on the chart for our northern winter. The lands tend to take tempera-

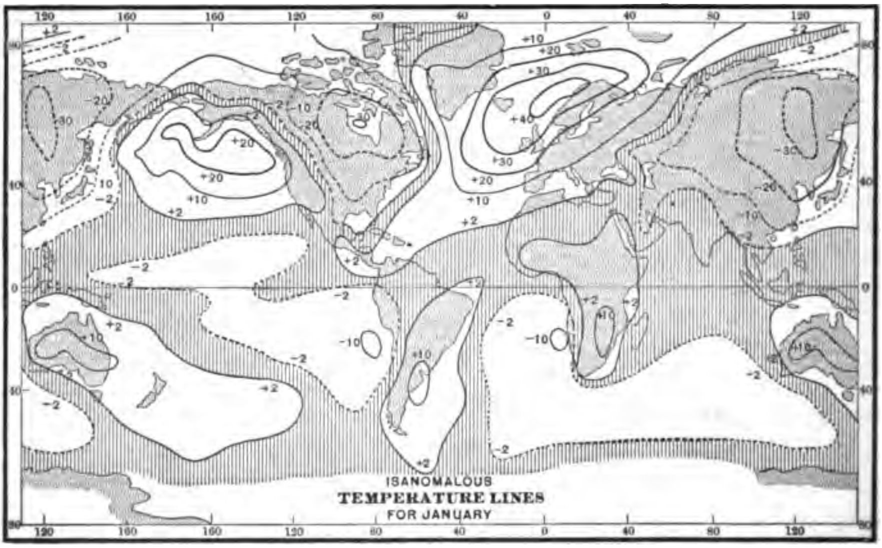


Fig. 16.

tures proportionate to their latitudes and proper to their season, while the waters try to maintain temperatures of the preceding summer, and the influx of equatorial waters greatly aids the northern seas in this attempt. The isotherms in our winter seasons are therefore extraordinarily deflected, particularly on the two sides of the North Atlantic; the isotherms are carried far to the south along our eastern coast, and far to the north along the western coast of Europe. In Lapland. the lines even lean over on their backs.

89. Thermal anomalies. The difference between the mean temperature of any place and the mean temperature of its latitude is called its thermal anomaly. The anomalies for January and July are illustrated in Figs. 16 and

17.[1] Areas not differing more than two degrees from the mean of their latitude are shaded with vertical lines.

For January the northeastern Atlantic and northwestern Europe are regions of excessive warmth for their latitude, being about 35° F. in excess of their normal. Northeastern Siberia is the region of the most excessive cold in the world, having a January temperature of 30° below its normal. The waters of the Alaskan Bay and the broken lands north of Hudson's Bay are the districts of too high and too low temperatures in the new world, their departures being about 20° above and 25° below their respective normals. Only small anomalies are found in the torrid and south temperate zones, as might have been foreseen from the regular course of their isotherms.

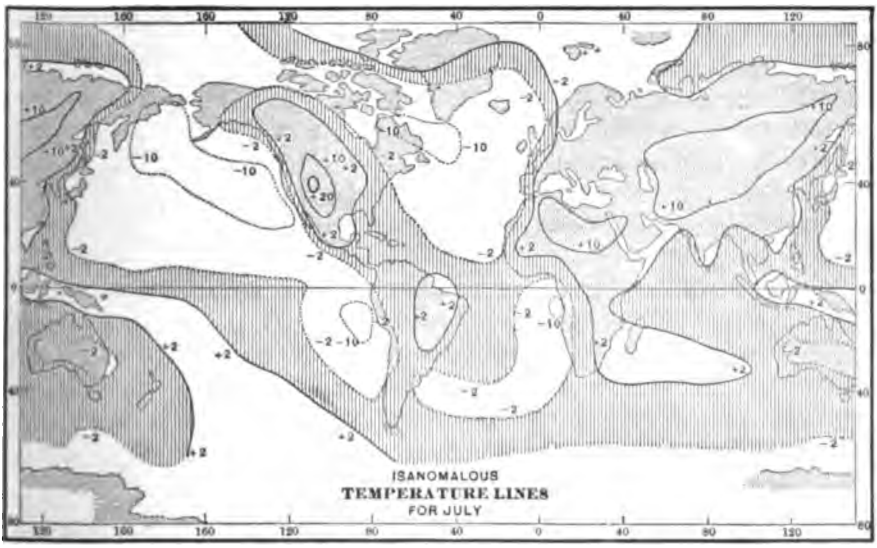


Fig. 17.

The July anomalies are less pronounced than those of January. The northern continents are warmer and the oceans are colder than the normals of their latitudes, but the departures are not so strong as before; excessive heat, 10° or more above the normal, is found from eastern Siberia to the Sahara and in our western interior deserts of Nevada and Arizona.

The annual anomalies exhibit positive and negative departures in the northern hemisphere similar to those of January, but less marked. A notable

[1] The charts from which these figures are reduced were constructed on the basis of Buchan's isothermal charts by Mr. S. F. Batchelder of the Senior class of Harvard College, 1893. The chart from which Fig. 18 is reduced was constructed by Mr. J. L. S. Connolly, of the same class. See American Meteorological Journal, vol. x.

negative annual anomaly is found near the equator west of South America, more strongly marked than anywhere else in the torrid zone; this being the result of the long reach made by the slender extremity of South America into the Antarctic eddy, thereby turning a great volume of cold southern water towards the equator. A peculiar effect of this cold current and the consequent anomalous temperatures that it causes on the equator is to exclude coral polyps from the Galapagos islands, while they flourish on the shores of similar islands further west in the Pacific, where the ocean current has become warm by longer exposure to equatorial sunshine.

60. Annual range of temperature. Several important generalizations are found on studying the distribution of large and small annual ranges of temperature, as determined by the difference between the means of the warmest and

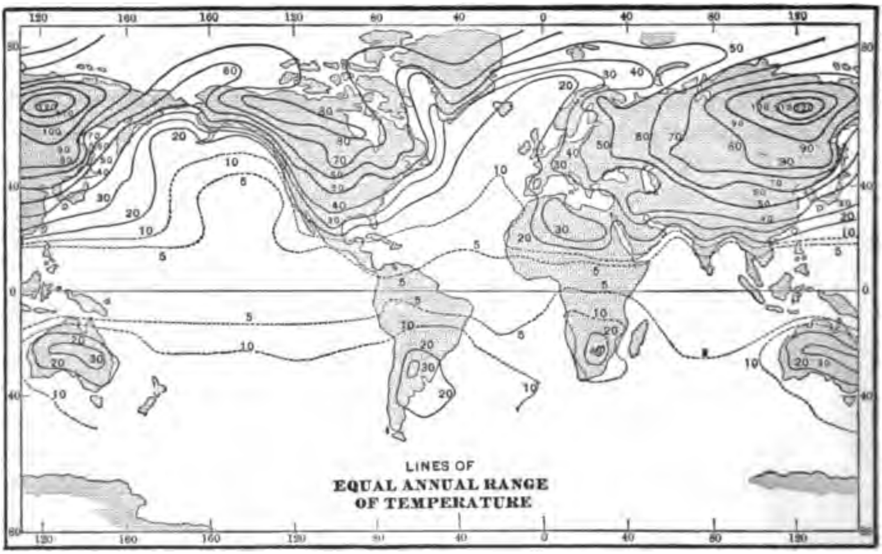


Fig. 18.

coldest months. Lines of equal annual range are drawn in Fig. 18.[1] In the first place, the area of moderate annual range — less than 10° F.— extends nearly all over the torrid zone, where the annual variation of insolation is small; and over a large part of the southern oceans, even to comparatively high latitudes. because the waters change their temperatures with so great difficulty; but on the polar seas, the annual range is strong in spite of their conservatism, because there the variation of insolation is so great. Passing next to areas of the most extreme range — over 70° — these will be found only on the large

[1] See note to Figs. 16 and 17.

land areas far from the equator; hence no areas of extreme range are found in the southern hemisphere; they are limited to the northern portions of the northern continents, where the strong variation of insolation readily causes a strong change in temperature. Great range of temperature, therefore, characterizes a continental climate in temperate latitudes, while a small range characterizes an oceanic climate in nearly all parts of the world.

A peculiarly unsymmetrical distribution of large and small ranges is found on the eastern and western coasts of our northern continents. A belt of small range of temperature — not over 25° — extends all along our Pacific coast, and all along the western coast of Europe; while the area of strong although not the most excessive range — over 45° — extends along our eastern coast and over the eastern coast of Asia. The reason for this is to be found in the combined action of ocean currents and winds, particularly in the control of the distribution of temperature by the latter. In temperate latitudes, the prevailing course of the winds is almost from west to east. The western coasts of the continents are therefore breathed upon by the winds coming from the oceans; these are comparatively mild in summer and cool in winter: hence the range of temperature that they allow over the coastal land is small. On the eastern coasts, the winds blow from land to sea and carry with them the extreme changes from cold winters to hot summers. The western coasts consequently savor of an oceanic climate; while the eastern coasts partake of a continental climate. The eastern coast of Asia is somewhat protected against the extreme cold of Siberia by mountain ranges some distance inland. The eastern part of the United States has no such barrier to keep back the cold winds from the far Northwest; hence the severity of our winter cold waves.

91. Polar temperatures. The table in Section 27 gave remarkably high values for the insolation received at the poles on the solstitial days when the sun rose to the greatest altitude over the polar horizons. If temperature followed insolation directly, we should find the hottest days of the world and of the year at the south pole on December 21, and at the north pole on June 20. As a matter of fact, the north and south poles are, as well as can be inferred, the coldest places in their respective hemispheres on these days. The reason for the small rise of temperature around the poles in spite of the great amount of insolation showered upon them is found, first, in the necessity of melting the ice and snow before the land temperature can rise above 32°; second, in the large water areas near the poles: third, in the brief duration of strong polar insolation.

Mention should be made of the peculiar migration of what is known as the northern "cold pole" or center of lowest temperature in the northern hemisphere in winter. In the southern hemisphere, we may reasonably expect the south pole to be the coldest spot throughout the year; for it is within an icy

plateau, surrounded by wide oceans. In the northern hemisphere, while it is true that for the mean temperature of the year, the region immediately around the north pole seems to be the coldest place in the hemisphere, and while the polar area is probably the coldest part of our hemisphere in summer also, this does not seem to be the case in winter, as far as the present records of Arctic temperatures go. In that season, in spite of the continuous darkness at the pole for five months, the temperature within the polar ocean is probably not below —42°, while in a certain part of northeastern Siberia, the mean temperature for January is —60°. The reason for this, and for many other facts of temperature distribution, is found in the more rapid cooling of land than of water, as well as in the circulation of the Arctic ocean waters, to which reference has already been made. The charts of our winter season therefore represent a strongly marked cold pole in northeastern Siberia, from which the temperature rises in all directions, north, south, east, and west.

It is possible, however, if future exploration discovers a considerable land area near the north pole, that the temperature there may fall to even a lower degree for January than is observed at the Siberian "cold pole." In this case we may speculate as to the seat of lowest temperature in the northern hemisphere in summer. A polar land area would reach a comparatively high temperature in June and July, if not too heavily clad with snow and ice; and it might then be even warmer than the ocean around it. In that case, the summer "cold pole" would be an annulus of low temperature around a polar oasis of somewhat higher temperature.

In later chapters, we shall return again to the question of temperature and its distribution in considering types of weather and the climatic features of the world; but before these are taken up, the control of the circulation of the atmosphere by its differences of temperature and the consequences of the circulation in producing clouds, storms and rain must be examined.

CHAPTER VI.

THE PRESSURE AND CIRCULATION OF THE ATMOSPHERE.

General Principles.

92. The conditions of general convectional motion. If the atmosphere were everywhere of uniform temperature, it would lie still on the earth's surface, and there would be no winds; but we have learned that the temperature of the atmosphere is continually or periodically higher in one region than in another, and that the chief variations in the distribution of temperature are systematically repeated, year after year. This prevents the stagnation of the atmosphere and ensures a systematic movement of air currents from place to place. The general principles on which such movements depend are extremely simple and will now be briefly stated, in order to prepare the way for a better understanding of the charts of atmospheric pressure and winds, which soon follow.

Let it be supposed that a certain part of the atmosphere is maintained at a higher temperature than that of its surroundings. The warmed air will be expanded; its upper layers will flow off to the surrounding regions, cooling as they go, and the pressure at sea-level will thereby be decreased in the warm region and increased round about it. The lower air, impelled by these differences of pressure, will creep in beneath the warmed air, warming as it goes, and thus a regular convectional circulation will be established on a large scale. As before, in Section 53, this may be likened to the working of a clock: we expend a certain amount of muscular energy in winding up the weight against gravity, and gravity then pulls it down again, driving the wheels and the hands. The sun warms a certain part of the air, thereby causing it to expand upwards against gravity; in other words, lifting the upper layers by the expansion of the lower ones. Gravity then pulls the upper layers down on the surrounding unexpanded air, and the differences of pressure thus introduced drive the other members of the circulation.

93. Arrangement of isobaric surfaces in a general convectional circulation. This problem deserves deliberate illustration. Let Fig. 19 represent a vertical section of the atmosphere, in which the vertical scale is much magnified compared to the horizontal. If the temperatures be uniform at every successive level, although decreasing at the usual rate from below upwards, then the isobaric surfaces will be level and concentric, as explained in Section 18. This arrangement is indicated by the broken lines, numbered at the margin to

indicate their respective pressures. Let it now be supposed that the temperature of all the central region is raised by a certain amount. All the air thus warmed will expand. The column *HA* will expand to height *HB*; and hence the isobaric surface of 29 inches will take the position *DBC*. All the overlying surfaces will be similarly bowed upwards, the highest of them being arched by the greatest amount. Consider now the condition of any two volumes of air at equal heights, one in the warmed region at *E*, the other in the cooler space at *F*. The pressure on *E* is 27.10; the pressure on *F* is 26.90. These volumes exert an expansive pressure in all directions equal to the pressure upon them. Hence the series of similar volumes between *E* and *F* will not be in equilibrium, but will be pushed outward to the left; and if this outward force suffices to overcome friction, movement in that direction will ensue.

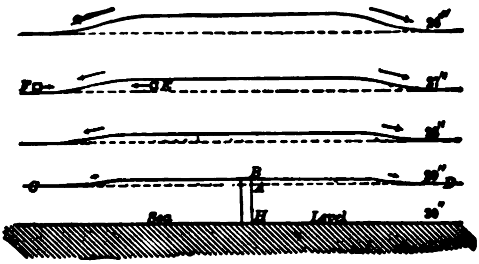


FIG. 19.

This condition of motion may be illustrated by the principle of the inclined plane, as follows. Let *ST*, Fig. 20, be a magnified part of one of the bowed isobaric surfaces of Fig. 19. As far as the air above this plane is concerned, all the lower air might for the moment be removed and replaced by any rigid body having a surface, *ST*. A volume of air, *G*, resting on the plane, is drawn down by the gravitative force, *Gg*. This may be analyzed into two components, one of which, *Gp*, at right angles to *ST*, causes no motion, while the other, *Ga*, parallel to the plane in the direction of its descent, urges the air to move down the slope.

In whichever way the problem is viewed, it is clear that the upper air above the warmed region is impelled to move away laterally and accumulate over the cooler surrounding region. In consequence of such movement, the pressures at sea-level will be rearranged. Let it be supposed that the pressure at *H* is thus decreased to 29.50; while that on the marginal area is increased to 30.25. What will be the position of the isobaric surface of 30.00 under this new arrangement? It could be found beneath *H* by descending a shaft about 450 feet deep, as at *J*, Fig. 21. It could be found above *K* by rising about 225 feet into the air. The pressure *H* being 29.50, and at *K* being 30.25, the pressure of 30.00 at sea-level may be expected at *M*, one third of the distance from *K* to *H*. The curved line *LMJM'L'* may now be drawn, as in-

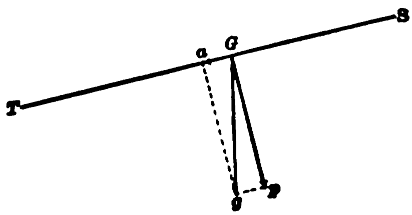


FIG. 20.

dicating with sufficient accuracy the new position of the isobaric surface of 30.00. From this as a base, the overlying surfaces may be constructed; for the height of a column of air between adjacent surfaces corresponding to a barometric inch under a given pressure and at a given temperature, may be taken from meteorological tables. It is, however, manifest from inspection that the distance between any pair of isobaric surfaces must increase in passing from the cooler to the warmer region; or, in other words, that the whole system of rearranged isobaric surfaces must diverge from one another as they enter the warm region. Hence the central depression seen in the concave surface of 30.00 will gradually weaken in the higher surfaces, and if we ascend high enough it must entirely disappear and be replaced by a central elevation in a series of convex surfaces. One of these is shown in the uppermost full line of Fig. 21. Between this line of 26 inches pressure and the isobar of 27 inches, other isobaric surfaces might be drawn for the fractional parts of the inch; and one of these at the height NN' would be level; this is called the *neutral plane*. Above it, the air tends to flow outward; below it, the air follows the slope of the isobaric surfaces and tends to creep inward.

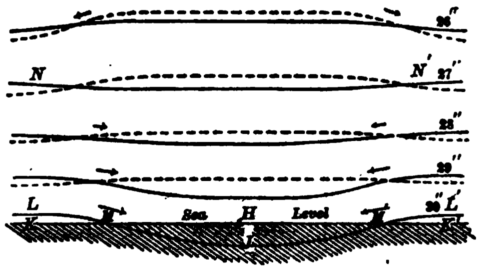


FIG. 21.

94. Conditions of steady motion. The arrangement of the isobars thus determined by the first outflow aloft suffers still further change on account of the inflow established below. The strong difference between central and marginal pressures that was first assumed is diminished, and finally falls to just that value by which a steady circulation can be maintained, as illustrated in Fig. 22. Inasmuch as the resistances to motion encountered by the air are very small, the final arrangement of the isobars has very faint slopes. The initial difference of temperature between the central and marginal regions is lessened by the interchange of air that it produces; but as long as any difference of temperature is maintained, a system of diverging isobaric surfaces will be maintained also, with outward slope above and inward slope below. The action of gravity on the inclined isobaric surfaces will then be entirely expended in overcoming the resistances

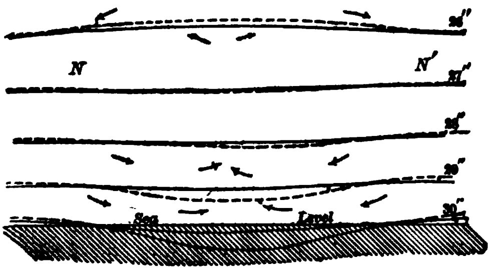


FIG. 22.

excited by the motion, and not in accelerating the motion to higher and higher velocities. This is like the case of a train of cars which an engine is pulling with all the force of high pressure steam, and which nevertheless does not exceed a certain speed of travel: the resistances excited by the movement of the train have then risen to equality with the pull of the engine, and no higher speed can be attained.

In the case of a large convectional circulation in the atmosphere, the velocity of steady motion will be greater if the central region is kept very warm. If the central region is maintained only a few degrees above the temperature of the surrounding region, the velocity gained will be moderate. If the supply of heat for the central region varies periodically, the differences of pressure produced and the velocities maintained by them will vary in the same period, changing with the rise and fall of central temperatures. If the central region is cooled below the temperature of the surrounding region, it will become an area of high pressure, with outflowing surface winds. If the central region is alternately warmer and colder than the surrounding region, the direction of the circulation will be changed in corresponding periods; the surface winds flowing inwards when high temperature and low pressure prevail, and outwards when the conditions are reversed.

95. Barometric gradients. The term, gradient, has already been introduced in connection with the vertical diminution of temperature, to indicate a rate of decrease. We now find need of the same term in connection with the decrease of barometric pressure in passing along a horizontal surface, such as sea level, from the margin to the center of a warmed region, or from the center to the margin of a cold region. If the slope of the isobaric surfaces is steep, the horizontal decrease of pressure will be rapid, and the gradient will be strong; if the slope is gentle, the gradient will be weak. Hence the barometric or baric gradient, commonly taken to measure only a rate of decrease of pressure along a horizontal surface, also indicates the amount of slope of an isobaric surface. The direction of decrease or of slope is commonly stated with the rate. The rate of decrease is commonly expressed in hundredths of an inch of pressure in a quarter of a latitude-degree of horizontal distance.[1] This subject will be met again in Section 113.

96. Vertical components of a convectional circulation. The ascent and descent of air currents caused by local diurnal convection have been described in Sections 45 to 54. The vertical movements of the air may become rapid under favorable conditions, as on warm level plains, when dust whirlwinds spring up towards noon. The vertical movement may then greatly exceed the

[1] This way of measuring the gradient is recommended because its numerical value is then unchanged if expressed in millimeters of pressure per latitude-degree of distance.

accompanying horizontal movements, both in velocity and in distance traversed; but in the examples of larger convectional circulations here considered, the case is quite different. The diagrams employed in the previous section are so greatly exaggerated vertically that they give wrong ideas in this respect, unless care is taken to conceive of the circulation in its true proportions. It must be borne in mind that all the larger examples of convectional circulation in the atmosphere have much greater horizontal than vertical dimensions. A vertical thickness of possibly twenty miles may be allowed for the general circulation between the equator and poles, and much less than that for the circulation between the continents and oceans; but the horizontal distances over which the circulating winds travel may be measured in hundreds or thousands of miles. Not only so; the cross-section of the ascending currents in the warm or cold central region may have a much greater area than the cross-section of the inflowing or outflowing winds; hence the velocity of ascent or descent in the central area may be small compared to the velocity of inflow or outflow around it. It follows from this that the vertical components of the larger atmospheric convectional motions are comparatively inconspicuous; they have low velocities, and the regions over which they occur occupy but a small share of the area swept over by the whole circulation; they may be much confused by local convectional currents, like eddies in the general downstream flow of a river.

It is desirable to gain a clear conception of these relations, as well as of the process by which the circulation of the atmosphere is kept up, in order to avoid certain careless forms of statement. It is not uncommon to hear it said: "The air is heated and rises, and the cold air rushes in from either side to fill the vacuum thus formed." It is better to express the facts of the case by saying: "As the air is heated, it expands and overflows aloft; the colder air then creeps in beneath from either side, warming as it goes, and raising the warmer air slowly above it." This places the slow ascensional movement in the warmed region at its true low value, and correctly suggests that the driving force of the circulation is found in the gravitative pressure of the colder air from the sides.

97. Application of the general principles of convectional motion to the case of the atmosphere. The knowledge gained in the foregoing chapter concerning the control, distribution and variation of temperature in the atmosphere should enable the student to make correct application of the principles stated in the preceding sections. He should expect to find a belt of low pressure around the heat equator, with caps of high pressure over the poles; the equatorial belt of low pressure should migrate north and south after the sun, and the contrasts between equatorial and polar pressures should be greater in the winter than in the summer hemisphere. The continents should have lower pressures than the surrounding oceans in summer, and higher

pressures in winter. The regions of marked abnormal temperatures for their latitude, such as those of abnormal warmth in the northern Atlantic and Pacific, should have correspondingly abnormal pressures. The winds should blow outward from the regions of high pressure towards those of low pressure; hence there should be a system of winds blowing from either pole towards the equator, but more or less modified by an indraft towards the continents in their summer season and an outflow in their winter. The uppermost currents should move opposite to the surface winds. The velocity of the winds should be greater where the barometric gradients are stronger; hence greater in the winter hemisphere as a whole. Along the axis of the equatorial belt of low pressure, as well as in the centers of the polar and continental areas of high or low pressure, where there are no gradients, there should be no winds; that is, calms should prevail. With these deductions in mind, the charts of pressure and of the winds for the year should be examined. Certain supplementary explanations must be introduced before all the facts concerning the pressure and circulation of the atmosphere can be understood, but no proper beginning can be made in this chapter without an understanding of the theory of convectional circulation.

The Measurement and Distribution of Atmospheric Pressure.

98. Measurement of atmospheric pressure. Reference was made in an earlier chapter to the pressure exerted by the quiet atmosphere upon the level surface of the ocean. The pressure would be uniform all over the world, if there were no differences of temperature and no winds. We must now investigate the means of determining the actual pressure at any place, and the general distribution of pressure over the surface of the earth under existing conditions. This requires, first, an examination of the construction and use of barometers, and, second, the study of charts on which the results of barometric observations are displayed.

Barometers are of two kinds. In instruments of one kind the pressure of the atmosphere is counterbalanced by the weight of a column of liquid of known density and measurable height; as the liquid employed is usually mercury, these are called *mercurial barometers*. In instruments of another kind the atmospheric pressure is balanced by a spring inside of a closed metallic box from which the air has been exhausted, so that any change in external pressure deforms the box slightly. These are called *aneroid barometers*, from being made without employing a liquid.

99. Mercurial barometers are made on the principle already explained on page 11. Various special devices are employed to simplify the measurement of the height to which the mercury column is held up in the tube above the

level of the surface of the mercury in the dish or vessel on which the air presses. The Fortin barometer, illustrated in Fig. 23, is most commonly employed in the stations of the Weather Bureau. The glass tube containing the mercury is enclosed in a brass tube, open in the upper part on two sides, so that the top of the mercury column may be seen; it is graduated to inches and tenths on the edge of the opening (*a*). The air gains access to the surface of the mercury in the vessel below through the fine crevices at the top of the vessel, where a glass ring (*b*) joins the base of the brass tube. The bottom of the vessel is a buckskin bag, within a brass cylinder (*c*), against which a thumb-screw (*d*) presses from below, so that the height of the mercury surface in the vessel can be raised or lowered until it just touches the end of a fine ivory pointer (inside the glass ring, *b*,) which represents the zero point of the brass scale. When thus set, the height of the mercury in the tube can be accurately read by a vernier or index (*v*) that slides in the opening of the brass tube. A good instrument of this kind costs about thirty dollars.

100. Correction for temperature. Readings thus made must be corrected for temperature, because of the expansion of the mercury as well as of the brass tube which carries the scale. If two barometers were under the same atmospheric pressure, one in the cold outer air and the other in a warm room, the latter would read higher than the former. To make allowance for this, the temperature of the barometer is determined by a thermometer (*t*) attached to the tube, and all readings are reduced to what they would be if the temperature of the whole instrument were 32°, by means of corrections given in barometric tables. This correction must be applied before different readings are compared.

101. Correction for latitude. The force of gravity, by which the spheroidal form of the ocean surface is determined, is not a constant, but varies from a maximum at the poles to a minimum at the equator, its greatest and least values being in the proportion of 193 to 192. It follows from this that if there were a uniform atmospheric pressure over the earth, the barometric readings at sea-level would vary; being progressively greater towards the equator and less towards

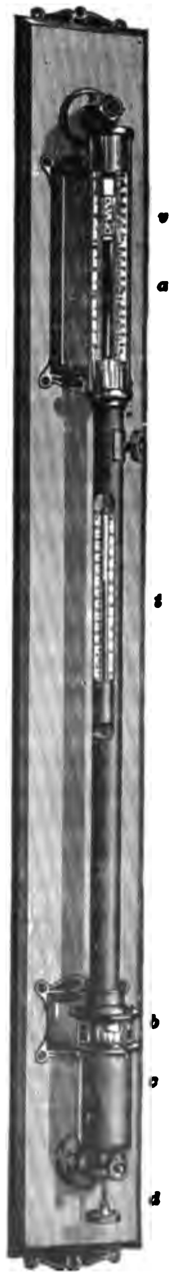


Fig. 23.

It follows therefore that this extremely small force is all that can be called on to move the air in the region referred to. The gradients in other regions may be somewhat weaker or stronger, but it will be found that in all cases only a small share of gravity is at work to maintain the circulation of the atmosphere.

HEIGHT OF A COLUMN OF AIR EQUAL TO $\frac{1}{10}$ INCH IN THE BAROMETER.

(Arranged from Hazen's Tables.)

PRESSURE.	0°	10°	20°	30°	40°	50°	60°	70°	80°	90°	100°
22" . . .	111	114	116	119	122	124	127	130	132	135	138
24 . . .	101	104	106	109	111	114	116	119	121	124	126
26 . . .	94	96	98	101	103	105	107	110	112	114	116
28 . . .	87	89	91	93	95	98	100	102	104	106	108
29 . . .	84	86	88	90	92	94	96	98	100	102	104
29.5 . . .	83	85	87	89	91	93	95	97	99	101	103
30.0 . . .	81	83	85	87	89	91	93	95	97	99	101
30.5 . . .	80	82	84	86	88	90	92	94	96	98	100

If these paragraphs have been appreciated, it is not too much to say that the chart of annual isobars will have by their assistance gained an entirely new meaning. The chart now represents not only a number of lines along which the pressure of the atmosphere at sea-level is equal; it represents the linear intersections of the sea-level surface with a system of warped isobaric surfaces, inclined one way or another at various angles. The gradient, which before represented only the direction and rate of decrease of pressure, or the inclination of the isobaric surfaces, is now given a definite value as a part of gravitative force. Where the adjacent isobaric lines are closer together on the chart, there the isobaric surfaces must be more steeply tilted, and hence there the winds may be expected to blow faster; where the lines are further apart, the inclination must be faint and the winds should be slow. Along the trough of the equatorial belt of low pressure, or along the axis of either tropical belt of high pressure, there must be strips of surface having practically uniform pressure; here the isobaric surfaces must be parallel to the surface of the sea; here the gradient must be zero; here calms should prevail, interrupted by light convectional breezes, and this we shall soon find to be the fact.

114. Isobars for January and July. Chart V, giving the mean pressures for January, differs from that of the year in several suggestive ways. The axis of the equatorial belt of low pressure — the barometric equator — is found to have shifted somewhat to the south of its average annual position; the northern tropical belt of high pressure has greatly broadened over the lands, where its pressure has increased, particularly over the greatest of all

is less in winter than in summer. In this country it is seldom over a tenth of an inch. At interior continental stations the subordinate oscillation diminishes in value. The curve for May in Fig. 25 is from a record at Harvard College, for a spell of fair spring weather, May 17 to 25, 1887, in which the diurnal variation is faintly shown; the maxima by +, the minima by O.

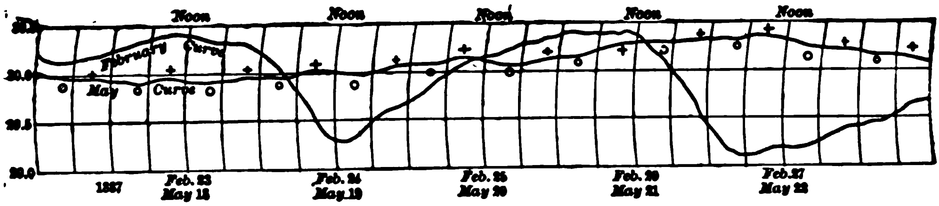


FIG. 25.

The cause of the diurnal variation of pressure is undoubtedly to be found in the diurnal variation of temperature, but the operation of the cause in producing the effect is not well understood.

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105. Irregular fluctuations of the barometer, having a period of one, two or three days, are caused by the alternate passage of areas of stormy and fair weather. Such changes are comparatively rare in the torrid zone; but they occur every three or four days in the greater part of the temperate zones, and in our winter season these fluctuations become so strong that the diurnal variation is hardly perceptible; as in the February curve in Fig. 25, copied from a barograph record at Harvard College for February 22 to 28, 1887, when two active stormy areas passed by, causing rapid weather changes. Fluctuations of this kind are further considered and illustrated in the chapters on storms and on weather. Much longer fluctuations have also been detected, covering several weeks or a month; these are generally called surges, but they have been little studied and their cause is not understood.

106. Barometer observations. Observations taken at 7 A.M., 2 and 9 P.M. will give a mean diurnal value of pressure practically free from the effects of the regular diurnal variation; but as the irregular fluctuations of pressure in our latitude are much greater than the regular variations, observations thus taken serve only to give the mean monthly pressure; and from these the mean annual pressure. Other values desired in discussing barometric observations are: the diurnal variation, determined from barographic records or from hourly observations; the mean monthly maximum and minimum, from which the mean monthly range is determined; the monthly extremes; the frequency of the irregular fluctuations and their average period and value.

107. Comparison of observations: reduction to sea-level. Barometric observations made at different heights above sea-level may be compared by their departures from their local mean annual or normal values; this being the method adopted in the early part of this century. A more satisfactory means of comparison is found in reducing all the observations to values that would have been obtained if they had been made at the same altitude; for example, at sea-level. This is done by adding to each reading (corrected for temperature) a supplementary pressure to make up for the imaginary column of air that may be conceived to reach downward from the station of observation to sea-level. The value of this supplementary pressure depends chiefly on three factors: first, the altitude of the station; second, the pressure at the station, for if the observed pressure be high, the imaginary column would contain air of greater density than usual; third, the temperature of the air, for if the temperature at the time of observation is higher than usual, the air of the imaginary column would be expanded to a comparatively low density. Tables for the reduction of barometric observations to sea-level, the altitude of the station being given, are furnished in the Instructions to Voluntary Observers, published by the Weather Bureau.

In preparing single barometric records for publication, all the data should be corrected for temperature; but it is best that they should not be corrected for altitude above sea-level. The corrections for reduction of the monthly means to sea-level should, however, be determined and published with the means themselves.

It must be remembered that the pressure indicated by the barometer corresponds to the weight of the atmosphere only when the air is calm. When it is moving, particularly when its velocity is high and variable, or when its temperature is rapidly changing, the atmospheric pressure on the barometer may be greater or less than the weight of the air; but the difference must be small in all cases; in thunderstorms, it may reach 1-20th inch (Section 254).

108. Barometric determination of altitudes. As the rate of the decrease of atmospheric pressure upwards is known, it follows that the barometer may be used in the determination of the altitude of a station above sea-level. This is commonly done in exploring expeditions and in preliminary surveys, the observations being reduced by specially prepared tables, such as those published by the Smithsonian Institution at Washington. The determination of altitudes used in reducing barometric observations to sea-level should, however, be made by careful levelling.

A convenient rule for finding the difference of level between two places by means of barometric observations is as follows: The difference of level in feet is equal to the difference of pressures in inches divided by their sum and multiplied by the number 55,761, when the mean of the air temperatures

at the two places is 60°. If the mean temperature is above 60°, the multiplier must be increased by 117 for every degree by which the mean exceeds 60°; if less than 60°, the multiplier must be decreased in the same way. For example, if the lower station has a pressure of 30″.00 and a temperature of 62°, and the upper station has 29″.00 and 58° respectively, the difference of level between the two will be

$$\frac{30-29}{30+29} \times 55{,}761 = 945 \text{ feet.}$$

If the lower values are 30″.15 and 65°; while the upper values are 28″.67 and 59°, then the formula becomes

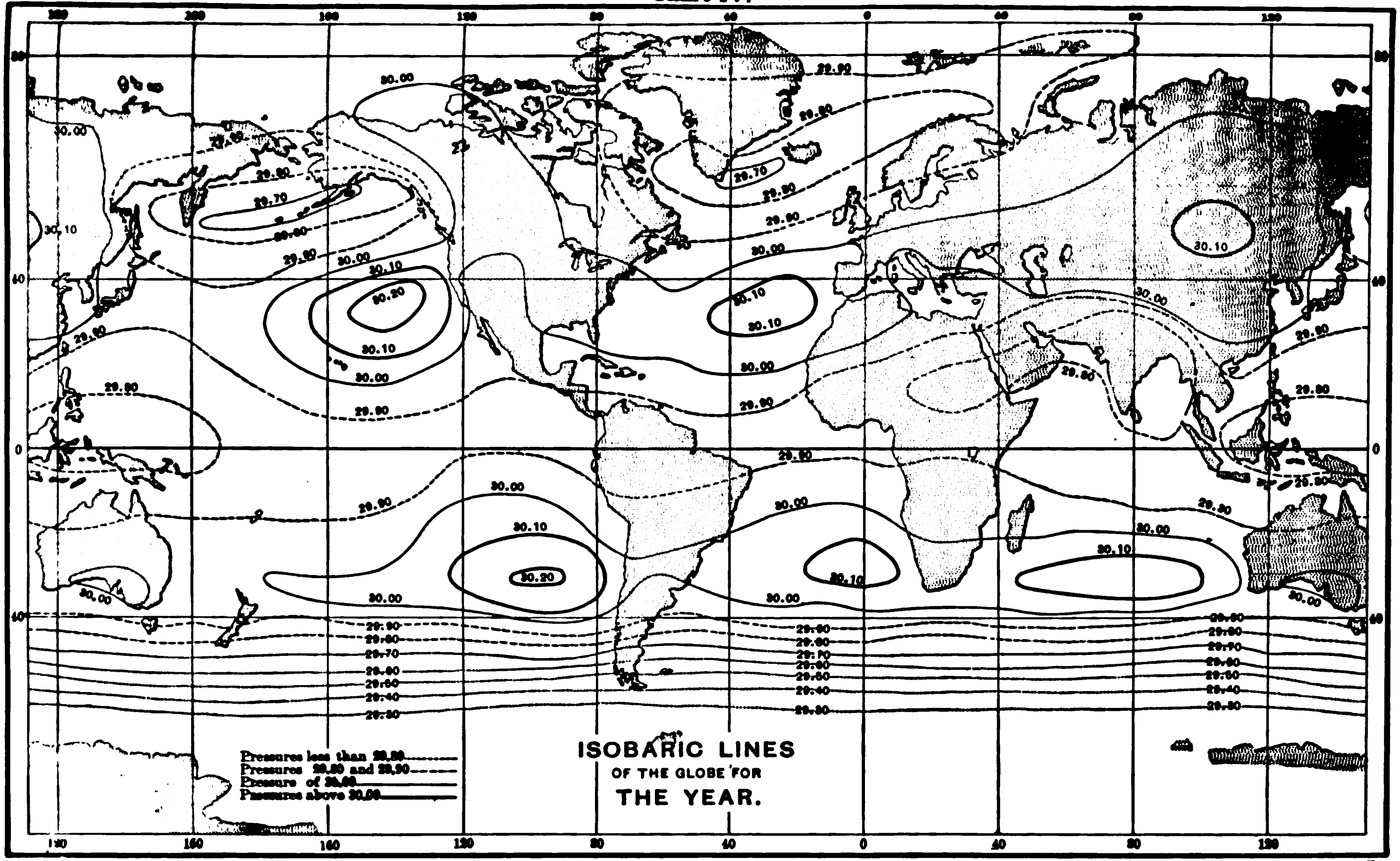
$$\frac{30.15-28.67}{30.15+28.67} \times [55{,}761 + (2 \times 117)] = 1409 \text{ feet.}$$

109. Barometric charts. Charts showing the distribution of atmospheric pressure are prepared in much the same manner as those already described for temperature. When observations have been continued for a number of years, the corrected mean annual and monthly values are reduced to sea-level and charted upon a map of the world; lines of equal pressure may then be drawn, these being called *isobaric lines*, or more briefly, *isobars*. The charts prepared by Buchan or by Hann are the most recent and complete (Section 80). Charts VI, VII, and VIII are reduced from those prepared by Buchan. The general distribution of pressure for the year and for January and July may now be considered.

110. Isobars for the year. The annual isobars on Chart IV show a belt of slightly diminished pressure running nearly around the equator; on either side there are belts of higher pressure, somewhat irregular in shape, with their middle lines about latitude 35° north and 30° south. These high pressure belts may be called the *meteorological tropics.*[1] The pressure then diminishes towards either pole, although in the northern hemisphere this diminution is much less marked and more irregular than in the southern; the lowest northern pressures being in the North Atlantic and North Pacific oceans. The differences of pressure thus found among the mean annual values are truly very small, their range from the highest in the North Pacific to the lowest in the far Antarctic ocean being only a little over an inch; but they are of great significance, as will be seen when the winds of the world are examined.

[1] The tropics are, etymologically, places of turning: hence the Tropics of Cancer and Capricorn, where the sun stops and turns in its annual migration north and south. The use of the term in the text here will be found later on to mark a limitation or turning in the course of the winds as well as in the direction of the gradients, of great climatic importance; greater, indeed, than that determined by the zenith altitude of the sun. Common usage often confounds the geographical tropics with the torrid zone which they bound. See footnote to Section 217.

Chart IV.



From Buchan's "Challenger" Report.

Bradley & Poates Engrs, N.Y.

111. Vertical section of the atmosphere along a meridian. If a vertical section of the atmosphere be drawn from pole to pole, it will be found to offer certain instructive contrasts to the condition described in Chapter II, where the isobaric surfaces of the atmosphere were imagined to lie level and essentially parallel and concentric under the action of gravity alone. Fig. 26 represents the desired section, the meridian line at sea-level being drawn as a straight line in order more easily to represent the attitude of the isobaric surfaces with respect to it. The vertical scale is greatly exaggerated. The poles are at N and S, the equator is passed at Q. The pressures at sea-level are taken from characteristic values on the chart of pressures for the year. The position of the isobaric surface of 30.00 is determined in the way employed in Section 93; here being seen as a line, it takes the shape $ABCDEF$. Other isobaric surfaces may then be added at greater heights, remembering that the lines by which they are represented must diverge in passing from the cold polar regions towards the warm torrid zone. The isobaric surfaces of 30.00 and 29.90 would be 98 feet apart at the equator, and 82 at the pole: the surfaces of 24.00 and 24.10 would be separated by about 116 and 96 feet, respectively. Consequently, the irregular warping of the lower isobaric surfaces is gradually changed to a system of regularly convex isobaric surfaces in the upper atmosphere. The equatorial belt of low pressure at sea-level has entirely disappeared at a height of about 12,000 feet; there is at that height and at all greater heights, a continuous poleward slope of the isobaric surfaces, of faint inclination in the torrid zone, but becoming much steeper in higher latitudes; the greatest inclination being in the southern hemisphere.

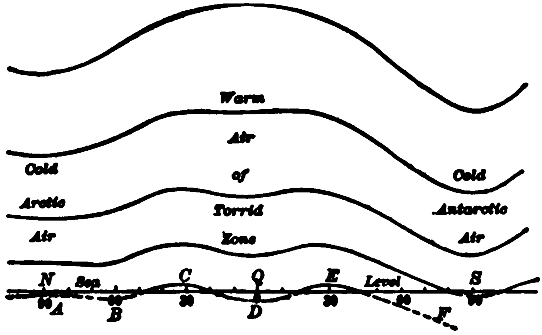


FIG. 26.

The arrangement of isobaric surfaces as thus determined is a matter of the greatest moment in considering the circulation of the atmosphere. Recalling what has been said about gradients, it is manifest that there must be a strong gravitative acceleration towards the pole in the upper atmosphere, and any theory of atmospheric circulation that does not take this fully into account is faulty.

112. Meaning of isobaric lines. The method of drawing isobaric lines on a chart has been briefly mentioned in Section 109; but it should now be perceived that every isobaric line on the charts represents the intersection

of some isobaric surface with the imaginary sea-level surface to which all barometric observations are reduced. Under the condition of uniformly distributed pressure, assumed in Section 18, there could be no isobaric lines, for the pressure at sea-level would everywhere be about 30.00; but as a matter of fact, the atmospheric pressure varies over the world; the isobaric surfaces are not level but are warped, as is shown in Fig. 26; and hence isobaric lines may be taken to indicate their intersections with sea-level. Given the isobaric lines on the chart, the isobaric surfaces which they stand for may be reconstructed. Any chart on which isobaric lines are drawn should be interpreted according to this suggestion.

113. Interpretation of gradients. Still another use may be made of Fig. 26. Suppose the section is drawn north and south through the middle of the Pacific, where the pressure at the equator is about 29.80, and where the isobar of 29.90 lies at latitude 10° N. Let this part of the section be drawn to a larger scale in Fig. 27, in which isobaric surfaces are represented for every

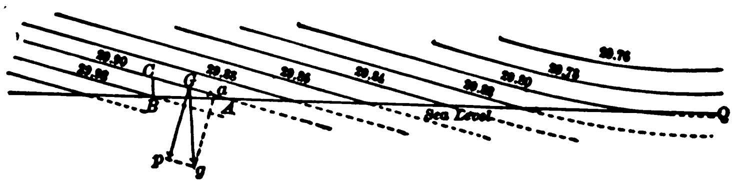


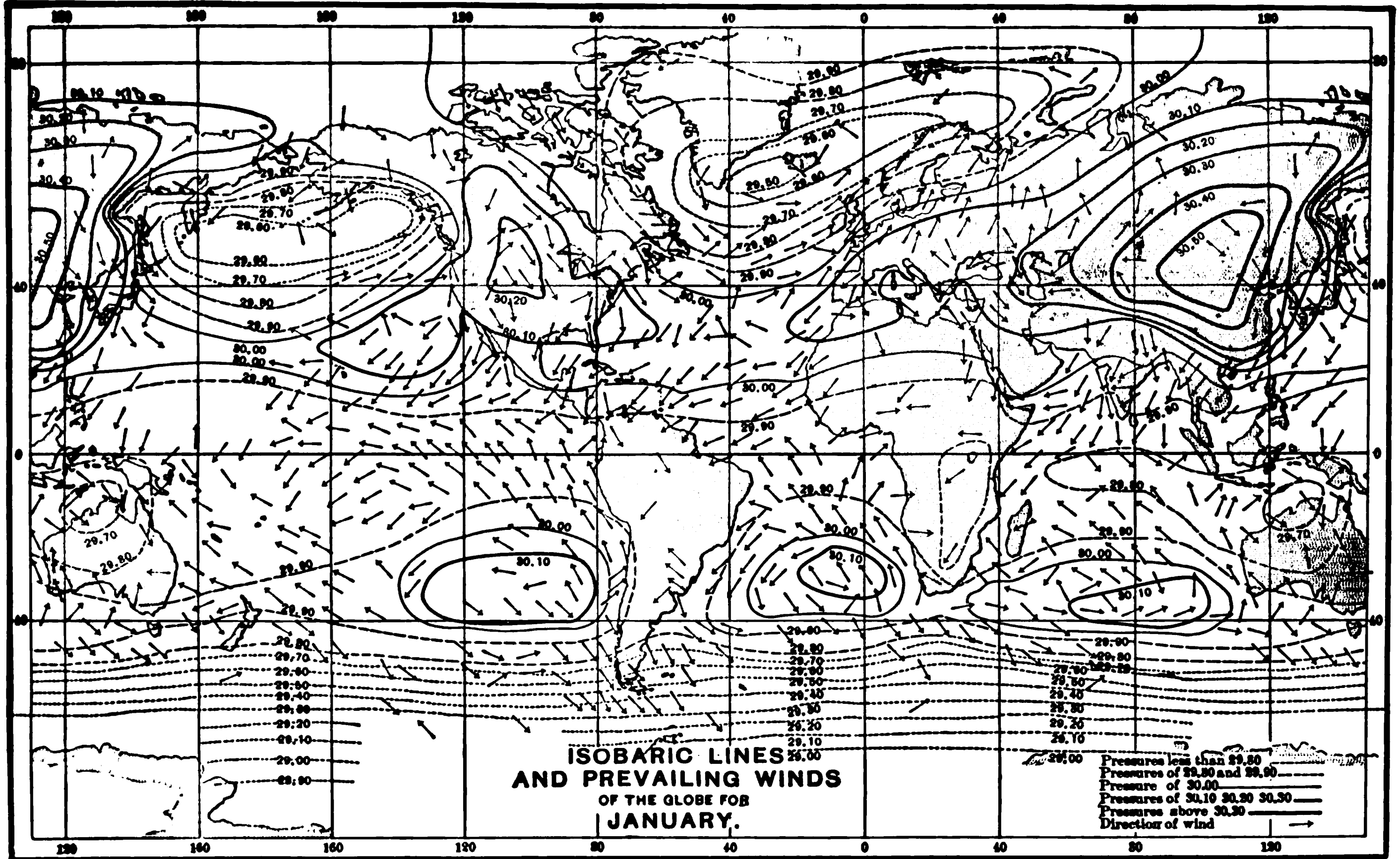
FIG. 27.

two hundredths of an inch. In this region it would be said: the barometric gradient is gentle, to the south. We may now proceed to calculate the value of the gradient in more definite terms. The action of gravity on an inclined isobaric surface, as explained in Section 93, must be recalled. By what small share of gravity will the air be urged to move equatorwards at the point A, where the isobaric surface of 29.90 intersects sea level? The dimensions of the triangle, ABC, may first be determined. The base, AB, is the distance along the meridian corresponding to an increase of 0.02 in pressure; and according to the chart of isobars for the year this is about two latitude degrees, or 140 miles, in the region under discussion. The height, BC, may be taken from the table given below, as equivalent to the height of a column of air under a pressure of 29.90 and at a temperature of 78°, and corresponding to two hundredths of an inch of barometric pressure. This is 19.4 feet. It is now desired to find the ratio of Ga to Gg. This may be done by the following proportion: —

$$Ga : Gg = BC : CA.$$

BA may be substituted for CA without significant error, and we have: The effective component of gravity at A is to the entire force of gravity as 19.4 is to 140 × 5280. The effective component is therefore only .000025 of gravity.

Chart V.



From Buchan's "Challenger" Report.

It follows therefore that this extremely small force is all that can be called on to move the air in the region referred to. The gradients in other regions may be somewhat weaker or stronger, but it will be found that in all cases only a small share of gravity is at work to maintain the circulation of the atmosphere.

HEIGHT OF A COLUMN OF AIR EQUAL TO $\frac{1}{10}$ INCH IN THE BAROMETER.

(Arranged from Hazen's Tables.)

PRESSURE.	0°	10°	20°	30°	40°	50°	60°	70°	80°	90°	100°
22" . . .	111	114	116	119	122	124	127	130	132	135	138
24 . . .	101	104	106	109	111	114	116	119	121	124	126
26 . . .	94	96	98	101	103	105	107	110	112	114	116
28 . . .	87	89	91	93	95	98	100	102	104	106	108
29 . . .	84	86	88	90	92	94	96	98	100	102	104
29.5 . . .	83	85	87	89	91	93	95	97	99	101	103
30.0 . . .	81	83	85	87	89	91	93	95	97	99	101
30.5 . . .	80	82	84	86	88	90	92	94	96	98	100

If these paragraphs have been appreciated, it is not too much to say that the chart of annual isobars will have by their assistance gained an entirely new meaning. The chart now represents not only a number of lines along which the pressure of the atmosphere at sea-level is equal; it represents the linear intersections of the sea-level surface with a system of warped isobaric surfaces, inclined one way or another at various angles. The gradient, which before represented only the direction and rate of decrease of pressure, or the inclination of the isobaric surfaces, is now given a definite value as a part of gravitative force. Where the adjacent isobaric lines are closer together on the chart, there the isobaric surfaces must be more steeply tilted, and hence there the winds may be expected to blow faster; where the lines are further apart, the inclination must be faint and the winds should be slow. Along the trough of the equatorial belt of low pressure, or along the axis of either tropical belt of high pressure, there must be strips of surface having practically uniform pressure; here the isobaric surfaces must be parallel to the surface of the sea; here the gradient must be zero; here calms should prevail, interrupted by light convectional breezes, and this we shall soon find to be the fact.

114. Isobars for January and July. Chart V, giving the mean pressures for January, differs from that of the year in several suggestive ways. The axis of the equatorial belt of low pressure — the barometric equator — is found to have shifted somewhat to the south of its average annual position; the northern tropical belt of high pressure has greatly broadened over the lands, where its pressure has increased, particularly over the greatest of all

land areas, the Eur-Asian continent;[1] the north polar pressure is somewhat higher than it was for the year, but the low pressure areas over the northern Atlantic and Pacific have become more strongly marked than before, and hence the general extra-tropical gradients of the northern hemisphere are stronger at this season than for the annual mean.

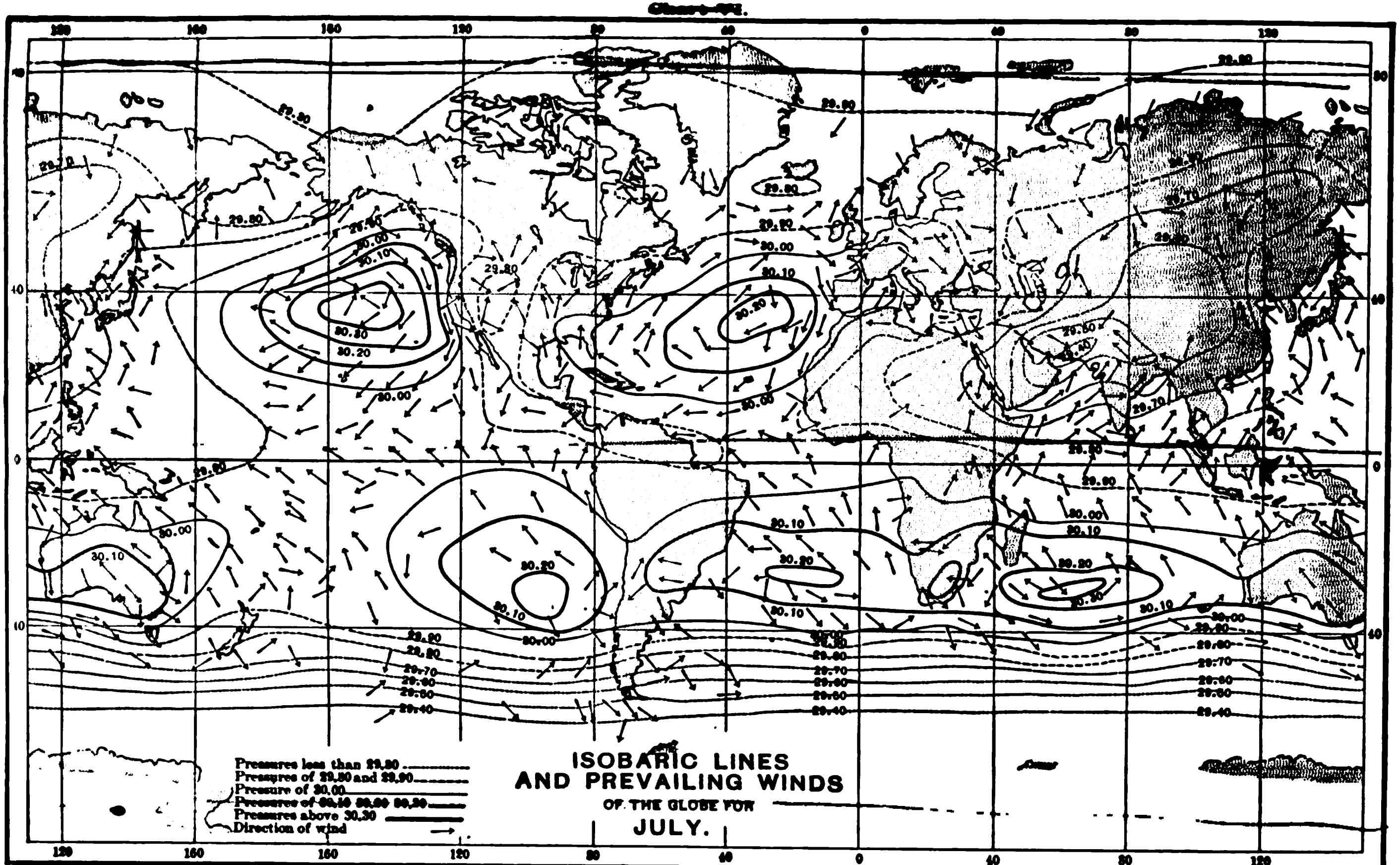
Passing now to the southern hemisphere, the tropical belt of high pressure is interrupted over the lands, and its average pressure is a little less than it was in the annual mean; the far southern latitudes show no significant change, observations not extending beyond latitude 70°, but the poleward gradients there are slightly weakened by the small decrease of pressure in the tropical belt.

In Chart VI, for July, the barometric equator has shifted to the north, and the shift is so strong over the lands that the northern belt of high pressure is destroyed, except on the ocean, where its remnants are correspondingly increased; they appear as two oval areas of distinct high pressure, which lie farther north than the axis of the belt in January; the north polar pressure is lower than in the chart for January, but the gradients around it are on the whole weaker.

In the southern hemisphere the high pressure belt is almost continuous around the world over land and sea, and its axis is farther north than in January; the polar pressures show no significant change as far as observed, but the poleward gradients are slightly stronger than before on account of a slight increase in the pressures of the tropical belt.

It has been calculated that the mean pressure over the whole earth for the year is 29.89 inches or 759.20 mm. The mean pressure for the northern hemisphere for January is 29.99; for July, 29.87: and for the southern hemisphere, 29.79 and 29.91. It appears from this that there is a much greater difference between the quantity of air over the two hemispheres in January than in July; and that the change is due to the shifting of an amount of air corresponding to a pressure of 0″.12 over a hemisphere, or a weight of about 32,000,000 tons of air, from one side of the equator to the other every half year. The relation of this semi-annual variation to the corresponding variation of temperature already considered is obvious.

[1] It should be understood that the charted increase of pressure in winter over continental plateaus is not an actual increase shown by direct observation, but only a relative increase seen after reduction to sea-level. The actual pressure on high plateaus is less in winter than in summer. The colder atmosphere in winter is compressed to lower levels, and leaves less air above the elevated plateau surface; but in the summer the expanded air from other parts of the world flows on the plateaus, and their pressure is increased. Yet if a comparison is made between the observed pressure on a plateau in winter and the calculated pressure at the same altitude over the adjoining ocean, the former will be found greater than the latter. It is this kind of comparison that is given on the isobaric charts, where all observations are reduced to the same level.



From Buchan's "Challenger" Report.

Bradley & Poates Engrs, N.Y.

115. Suggested explanation for the distribution of pressure. If it were not for the low pressures at the poles, one might say at once that the distribution of pressure is determined by the distribution of temperature. Where high temperature prevails, the air would expand; its upper layers would flow off, leaving low pressure, and accumulating in regions of prevailingly low temperature. Areas of high and low pressure should therefore be expected in areas of low and high temperature. This is indeed true to a certain extent: there is a belt of low pressure around the heat equator; and this low pressure belt migrates north and south with, or a little after, the heat equator. The continents have low pressure in summer and high pressure in winter, as their thermal relations to the surrounding oceans would have led us to expect. The areas of abnormally high winter temperature far north on the Atlantic and Pacific oceans have distinctly low pressure at the same time. But no cause is apparent for the tropical belts of high pressure; and at the poles, where the low temperature of winter in particular would lead us to look for very high pressures, the facts contradict our expectation in the most emphatic manner. The polar regions have relatively low pressure the year round: the whole south frigid zone has lower mean pressures than occur anywhere else in the world.

Before attempting further search for an explanation of the distribution of pressures, the general circulation of the winds must be examined.

Observation and Distribution of the Winds.

116. Winds. Air moving near the surface of the earth and in a nearly horizontal direction is called *wind*. Other motions may be called air currents; but they also are often called winds in a general way, as the "upper winds."

Observations of the wind should include its direction and its force or velocity. The direction is determined by a wind vane, moving freely on a vertical axis on some elevated spire or pole. It is always to be recorded as the point of the horizon from which the wind comes. The arms, bearing the letters, N, E, S, W, by which the vane is read, should be carefully set to the four cardinal points, allowance being made for the local variation of the magnetic needle from the true north. Intermediate points should be recorded as NW, ENE, etc. The direction from which the wind comes is called windward; that to which it goes, leeward. A change in the direction of the wind is called veering when it progresses from left to right; and backing when the shift is the other way. The amount of change is often expressed in points, a nautical term meaning an eighth of a quadrant, or 11¼ degrees.

117. The anemoscope gives an automatic record of the direction of the wind. The wind vane turns a vertical rod that reaches down into a room

below: a cylinder is attached to the lower end of the rod and turns with it: a pen presses lightly on a paper wrapped around the cylinder: the pen is carried downward at a slow and regular rate by clock-work, so as to descend through the length of the cylinder in a day.

The vertical component of the wind's motions is not detected by ordinary anemoscopes. An anemoscope may be specially arranged for this purpose, having a vane that moves on a horizontal axis, and which is always pointed towards the wind by a larger vane moving on a vertical axis; but this is seldom used.

118. Force and velocity of the wind. The force of the wind may be estimated or measured. The following scale is recommended in rating different velocities.

TABLE — *Scale, velocity and pressure of winds.*[1]

		AVERAGE VELOCITIES.		AVERAGE PRESSURES.	
SCALE.	TERMS.	Miles per hour.	Meters per second.	Pounds per square foot.	Kilograms per square meter.
0	Calm	0	0	0	0
1	Very light breeze	2	1	0.03	0.15
2	Gentle breeze	7 or less	3 or less	0.23 or less	1.13 or less
3	Fresh breeze	11	5	0.64	3.15
4	Strong wind	18 or more	8 or more	1.62 or more	7.97 or more
5	High wind	27	12	3.64	17.9
6	Gale	36	16	6.48	31.9
7	Strong gale	45	20	10.12	49.8
8	Violent gale	58	26	17.12	84.2
9	Hurricane	76	34	29.26	143.9
10	Most violent hurricane	95	42	45.12	222.0

A scale of six numbers is often used, its terms being light wind, moderate wind, strong wind, fresh gale, whole gale, hurricane. It is generally the case that the higher numbers of these scales are too frequently employed.

119. Anemometers. It is apparent that estimates of the velocity or force of the wind must be very faulty; and that instrumental records are much to be preferred. A simple indication of the pressure of the wind, from which the velocity may be obtained by the above table, is obtained by means of Lind's wind-gauge: this consists of a tube, bent in the form of U, containing a liquid in its lower curve. One end of the tube, bent horizontally, is always directed to the wind by a vane. The pressure of the wind on the liquid then raises the surface of the liquid in the further arm of the tube, where its height may be

[1] The indefinite values of Nos. 2 and 4 result from the attempt to express the Smithsonian or Voluntary Observer and the International Bulletin scale in a single table.

read on a scale. This instrument has the value of being at least consistent in different places and at different times; while estimates of wind force by different observers cannot be closely comparable.

Another simple instrument for determining the force of the wind consists of a square board or sheet of metal, hinged along its upper edge, and always turned to face the wind by means of a vane. The deflection of the square from a vertical position gives a measure of the violence of the wind. The maximum deflection may easily be automatically registered, thus giving indication of the highest force reached by the wind since the last record. Sometimes the board is fixed in a vertical plane, but is moved by the wind against a spring; this may also register its maximum record.

The instruments just described can hardly be trusted to give good measures of the velocity of the wind. For this purpose, more accurate methods must be adopted, as in the Robinson anemometer, Fig. 28. Four arms are fixed at right angles, carrying hemispherical cups at the ends and rotating on a vertical axis. It is found by experiment that the wind moves with about two and a third or two and a half times the velocity of the cups. The vertical axis may be connected with a train of cog wheels, so arranged as to count the rotations of the arms and register the number of miles run by the wind on a dial that may be read at certain hours; or it may make a continuous record on a cylinder rotating by clock work, on which a pen marks the successive miles of wind, generally by means of some electrical device, or in some other manner, and the instrument is then called an anemograph.

FIG. 28.

When the wind is strong, the momentum of the whirling cups carries on their movement through temporary lulls or slatches of the wind, and hence at such times the recorded number of miles traveled is too great.

The velocity of the wind in miles per hour may be reduced to the velocity in meters per second (usually employed in Europe) by multiplying by 0.447; meters per second may be reduced to miles per hour by multiplying by 2.237 or about 2¼.

An anemograph for vertical currents has sometimes been employed, using a helical fan on a vertical axis to measure the upward or downward movement of the air. The motion thus determined is a small part of that taking place in a horizontal direction; it increases at times of convectional movements.

120. Hill and mountain observatories. The observations described above suffice to determine the characteristic movement of the wind about the place of observation; but in studying the general movements of the atmosphere, it is desirable to exclude the local shifts of the surface winds, and for this purpose, observations on hills or lofty towers are very useful. The observatories established by Mr. A. L. Rotch, on Blue Hill, near Boston, 635 feet above the sea, and by Mr. W. L. Childs on Mt. Wantastiquet, N. H., opposite Brattleboro, Vt., at a height of 1076 feet above the adjacent valley and 1364 feet above sea-level, are of much interest in this respect. The Eiffel tower, 990 feet high, erected in Paris for the Exposition of 1889, gave a series of particularly instructive meteorological records, inasmuch as its slender form produced practically no disturbance in the conditions of the air around it; while there is reason to think that hills and mountains, projecting into the atmosphere in large mass, cause a somewhat more hurried flow of the wind over their summits than is found from the velocities of floating clouds at the same height in the free air. Yet it is to mountain observatories at great heights that we owe at present the most definite information concerning the movements and other physical features of the upper strata of the atmosphere. Records from balloons are temporary; records from clouds refer only to direction and velocity of movement, and are moreover not obtainable in clear or in heavily clouded weather. Records on mountains may be maintained continuously and completely, although involving great expense. The mountain observatories established by our Signal Service, the official predecessor of our present Weather Bureau, on Mt. Washington, N. H., in 1871, at an altitude of 6279 feet, and on Pikes Peak, Colo., in 1873, at an altitude of 14,134 feet, have yielded results of great scientific value; but on account of their heavy expense and their relatively small value in the daily work of the Service in predicting the weather, these were discontinued in 1887 and 1888. The Lick Observatory, on Mt. Hamilton, Cal., 4400 feet above sea-level, maintains a full meteorological record. Automatic records are maintained by the Harvard College Observatory, on Mt. Chachani, Peru, at an altitude of 16,650 feet.

The most important of the mountain observatories in Europe are named in the following table.

Peak.	Mountain Range.	Country.	Height in Feet.
Ben Nevis	. . . .	Scotland	4407
Brocken	Hartz	N. Germany	3743
Schneekoppe	Riesen Gebirge	Germany	5246
Wendelstein	Alps	Bavaria	5669
Hoch Obir	E. Alps	Austria	7047
Sonnblick	E. Alps	Austria	10,155
Sentis	Alps	Switzerland	8215
Puy de Dôme	. . . .	France	4800
Pic du Midi	Pyrenees	France	9381

121. Wind observations. It is customary to make observations of the wind at the hours selected for other observations, as of temperature or pressure. It should be noted that the surroundings of an observer have a strong influence on the accuracy of the wind-record. In a city, the wind is continually thrown into irregular gusts and whirls by the many uneven obstructions that it must pass over; in a valley, the velocity is reduced and the direction is altered by the protecting hillsides. On an open prairie, there is good opportunity of securing comparable results; but even then, unless the vanes and anemometers of different observers are placed at the same height above the ground, the results are not closely accordant; for the wind is always much retarded by the resistances felt near the ground, and its velocity decreases rapidly as one approaches the surface of the land. At sea, the velocity of the wind is much greater than on the continents, and it is probable that the increase in the velocity with height is much slower than on land.

No standard height for vanes and anemometers has yet been adopted, because it is generally impossible to conform to any prescribed rule in this respect; but a height of at least 70 feet above the open ground is strongly recommended.

The increase of the velocity of the wind at considerable elevations has been determined by observations of clouds (Section 212); the results of such measurements at Blue Hill Observatory, near Boston, Mass., may be introduced in this connection.

Mean Cloud Velocities at various Altitudes: Blue Hill Observatory.

Altitude, meters		200 to 1000	1000 to 3000	3000 to 5000	5000 to 7000	7000 to 9000	9000 to 11,000	11,000 to 13,000
Mean velocity in met. per sec.	Summer	7.5	8.2	10.6	19.1	23.5	31.1	35.2
	Winter	8.8	14.7	21.6	49.3	54.0	. .	. .

122. Reduction of observations. Wind observations are commonly reduced by counting the number of times the various directions are recorded, and averaging these and the corresponding velocities with respect to the hour, the month and the year. The percentage of calms in the total number of observations should be determined. More elaborate reductions require an analysis of directions and velocities, so that the resultant movement of the air may be determined; but this is not a useful method where the direction changes frequently and irregularly, as with us.

In illustration of hourly observations of the wind, reference may be made to Fig. 44, showing the average direction of the wind for every hour of the day during the month of July, 1882, at the Lake Crib, or tower from which the water supply was taken for the city of Chicago from Lake Michigan; the regular veering of the lake breeze by day into the land breeze by night is thus exhibited to a nicety.

The average annual frequency of the winds at Kinderhook in the Hudson valley, trending north and south, and at Utica in the Mohawk valley trending east and west, both in the State of New York and a little over a hundred miles apart, are graphically presented in Figs. 29 and 30; this illustrates very clearly the control exerted by even these broad valleys on the course of the prevailing winds.

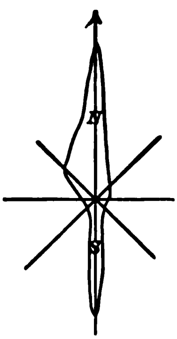


FIG. 29.

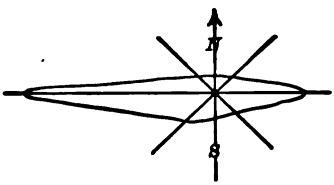


FIG. 30.

In the monthly reports from an observing station, it is customary to give the prevalent direction or directions of the wind, and the total wind movement for that time; thus for January, 1885, Pikes Peak, Colo., altitude 14,134 feet, had winds prevailingly from the northwest, with a total movement of 13,816 miles; Sandy Hook, at the entrance to New York harbor, with an anemometer close to sea-level, also had winds generally from the northwest, with a total movement of 14,932 miles. The example is taken from the Monthly Weather Review of the Weather Service, and is interesting in showing an exceptional greater activity of the surface winds than of the upper currents for the month in question; but if the resultant of all the movements of the wind were taken, it is probable that the back and forward winds of Sandy Hook would exhibit a smaller general progression of the air in any single direction, while on Pikes Peak nearly all the movement was in one direction, showing a great forward movement of the atmosphere over the mountain summit.

Ocean charts for the use of mariners have been prepared by the Hydrographic Offices of different countries, showing among other data the relative frequency and average force of the winds from the different directions during the several months and for the separate latitude and longitude "squares" of the ocean areas. The parts of the oceans frequented by numerous vessels are well explored in this respect; but there are still large areas where the knowledge of maritime meteorology is deficient.

123. The general winds of the world are best studied on the seasonal charts of pressure (Charts V and VI). As the winds of the southern or austral hemisphere are more easily generalized than those of the northern or boreal, they will be most frequently referred to in this general account. A fuller account of the winds of different regions will be given in the next chapter.

In July the winds between the southern tropical belt of high pressure and the equatorial belt of low pressure blow steadily from the southeast; these are the southern members of the *trade winds*. In the great circumpolar area beyond the high-pressure belt, the winds blow briskly from the northwest or west-northwest, but are much confused by stormy interruptions; these will be called the *prevailing westerly winds*. In the middle latitudes of the south temperate zone the winds are so boisterous as to have gained the name of the "roaring forties." The directions here named are much affected on and in the neighborhood of the land; but as the austral continents are relatively small, the winds of that half of the world blow for the most part as has been briefly stated. In January the division between the westerlies and the trades is less distinct than in July; in the South Atlantic, for example, when the continuity of the tropical high-pressure belt is broken, the winds seem to circulate around a central area of high pressure; while in July, when the tropical high-pressure belt stretches across land and water, it separates the two members of the wind system by a more linear division. In July, moreover, the austral winds on the whole blow faster than in January.

In the northern hemisphere the winds are in a general way symmetrical with those of the southern. There is a northeast trade wind over the zone from the tropical high-pressure belt to the belt of low pressure around the equator; and on the oceans at least there is a prevalent west-southwest wind over a considerable part of the area from the tropical belt towards the pole, with higher velocities in January than in July; in latitudes above 60° N. northeast winds are frequently recorded. But as the lands here occupy so much greater a share of the total area than in the southern hemisphere, the irregular winds that they produce greatly distort the boreal wind system. The simplest statement of this distortion is that the winds tend to blow obliquely outward from the continents in January, and obliquely inward in

July; but this tendency is greatly modified by the presence of the prevailing westerly winds in the latitudes where the continents are broadest. A similar effect is shown in the seasonal variation of the winds over Australia.

The axes of the tropical high-pressure belts and of the equatorial low-pressure belt, where the gradients are zero, are characterized, especially on the oceans, by weak and variable winds with not infrequent calms, in strong contrast to the continuous movement of the steady trades on the one hand, or to the stormy westerly winds on the other. The equatorial belt in particular is marked by frequent calms, called the *doldrums*, which migrate north and south with the ~~barometric equator, which in its turn follows the~~ heat equator; but it is an exaggeration to describe the doldrums as a belt of continuous calms. The light winds and calms of the tropical belts mark the "horse latitudes," and these also have a slight annual migration north and south with the sun.

Comparison of the Consequences of the Convectional Theory with the Facts of Pressure and Winds.

124. General relation of winds and pressures. The winds show a distinct tendency to blow from areas of high pressure towards areas of low pressure; they blow faster on the steeper gradients of winter than on the fainter ones of summer; they are boisterous on the steep gradients of the south temperate zone, and they weaken almost to stagnation where the gradients disappear along the axes of the pressure belts. According to the general principles of convectional circulation, as stated in Section 93, it was expected that the winds should follow the line of the gradient; but this is clearly not the fact. The boreal winds turn to the right; the austral winds turn to the left of the gradient.

125. Agreements and disagreements. This general review of the distribution of pressures and circulation of the winds has discovered two particulars in which the expected arrangement of pressures and motions are contradicted by the facts. The polar pressures are not high, but low; and the pressure is highest around the tropics, where intermediate values were expected: the winds do not flow along the gradients, but turn systematically to one side or the other. Otherwise, the consequences of the convectional theory accord with the facts.

When an investigation reaches this stage, the student may review its progress in some such way as this: Either the suggested explanation by means of convection is fundamentally wrong, in which case it should be replaced by another; or the explanation needs some supplements by which to account for the polar low pressures and the oblique course of the winds. It

can hardly be supposed that an explanation as well grounded on accepted physical laws as the one outlined in the statement of the general principles of convectional motion should be entirely wrong; moreover, such facts as the seasonal variations of pressure and winds over the continents and the frequency of calms along the axes of the pressure belts on the oceans are much in its favor. It should not be discarded until the possibility of supplementing its deficiencies has been very carefully tested.

Next it might be asked whether each one of the two classes of deficiencies requires special supplementary explanation; or whether one class may not be related to the other as a cause is to an effect; for in that case a single supplement that would explain the first would explain the second also. Let us consider if the oblique course of the winds can account for the unexplained distribution of pressure at the poles and the tropics.

126. Low polar pressure caused by the prevailing westerly winds. Brief mention was made in Section 13 of the flattening of the earth at the poles by the centrifugal force of its diurnal rotation, once in twenty-four hours. If it should rotate faster, it would be more flattened; if slower, it would be more nearly spherical. Look now at the winds of the southern hemisphere, from the tropical belt of high pressure to the pole. They are moving eastward over the earth's surface in a great whirl around the south pole. The upper winds, carrying the clouds, run even faster eastward than the surface winds. As a whole, they accomplish a revolution around the earth's axis in less than twenty-four hours; and hence their centrifugal force must be greater than that of the earth. May it not be that the expected high pressure at the pole is reduced to low pressure by the excessive centrifugal force of the circumpolar whirl, and that the air thus withheld from the polar region is found in the tropical belt of high pressure? This suggestion is certainly too important to be neglected; it is plausible enough to warrant its provisional acceptance, while search is made for the cause of the deflection of the winds from the gradients, whereby the circumpolar whirl is produced.

The Effects of the Earth's Rotation.

127. The deflecting force of the earth's rotation. The cause of the deflection of the winds from the gradients is to be found in the earth's rotation. It may be easily explained and illustrated by experiment (see Sect. 133) that the winds cannot follow the gradients, because there arises from the earth's rotation a force[1] that tends to deflect all horizontal motions, of what-

[1] Although always spoken of as a "force," this term implies a misconception of the same kind as that which often embarrasses the understanding of "centrifugal force." A body moving without friction over the surface of the earth tends to move in the direction of its

ever direction, to the right in the northern hemisphere, and to the left in the southern: the deflecting force is proportionate to the velocity of motion, and increases from zero at the equator ~~to a maximum~~ value at either pole.

The value of the deflecting force, in terms of the weight of the moving body, may be determined for any latitude by multiplying the appropriate factor in the following table by the velocity of motion, expressed in miles per hour.

LATITUDE.	FACTOR.	LATITUDE.	FACTOR.
0°	0.00,000,000	50°	0.00,000,509
10°	115	60°	576
20°	228	70°	625
30°	333	80°	655
40°	427	90°	665

128. Hadley's theory of the effect of the earth's rotation: 1735. The introduction of this important principle into our science has been slow, and even to this day it is not properly appreciated by many students of the subject. The oblique movement of the trade-winds was known, from the accounts of navigators, to Halley, a famous English astronomer, who tried to explain it in 1686 as a result of the (apparent) westward movement of the sun around the earth. In 1735 this was shown to be wrong by Hadley, another English astronomer, who introduced the first reference to the real cause; but his essay was generally overlooked for the greater part of the last century, until the same idea had occurred to several other investigators. The explanation still generally current follows that given by Hadley, in brief as follows: If a mass of air moves from latitude 30° north towards the rarefied belt of heated air around the equator, it advances upon latitudes whose eastward rotary velocity is greater and greater, and in consequence of this, the air lags behind, and hence appears as an oblique northeast wind; indeed, if it were not for the friction with the surface of the land and sea, by which the advancing air continually acquires something of the eastward motion of the latitudes that it enters, there should be a violent westward hurricane, of a hundred or more miles an hour at the equator, according to this theory. Hadley did not apply

first impulse. We live on the earth's surface, unconscious of its rotary movement, and consequently persuaded that any straight line holds a fixed direction. Hence, when a free-moving body (such as a free-swinging pendulum, as in Foucault's experiment) turns aside from its first line of movement, we assume that its direction has been changed by some deflecting force. In reality, the free-moving body perseveres in its original direction, in virtue of its inertia; it is the apparently fixed line of reference that is changing its direction, in virtue of the earth's rotation. The "deflecting force" is therefore only the inertia-resistance that a free-moving body exerts against a constraining force that urges it to move in what we call a straight line or a fixed direction.

his explanation to the prevailing westerly winds; but it has been applied to them by his followers, who teach that as these winds advance northward from the tropical belt, they enter latitudes whose eastward velocity is less and less; and in consequence of this the winds run ahead of the surface and gain a direction from the southwest or west-southwest.

This explanation contains two serious errors, which are here referred to because they have gained general currency. Hadley's explanation implies that there is no effect produced on motions to the east or west, while as stated above, the deflective force arising from the earth's rotation is independent of the direction of motion. Again, Hadley's explanation teaches that a body moving towards the equator continually lags *westward*, so that if friction had no effect it would attain a great velocity to the west when it reached the equator. This is wrong; the lagging, if such an expression is introduced at all, cannot be continually in one direction, as to the west, but can only be at *right angles to the momentary direction of motion*, and hence can produce *no effect on the velocity*. If a body were given a velocity of 25 miles an hour to the south when in latitude 30° N., and was supposed to move without friction over a level surface, it would continue to move at the same moderate rate whatever latitude it reached; while Hadley's explanation would give it a velocity of a hundred or more miles westward at the equator. A proper understanding of the true value and action of the deflective force should therefore be introduced into the popular teaching of meteorology.

The deflective effect of the earth's rotation was worked out by various mathematicians in the early part of this century; but its first proper application to meteorology was in 1843, when Charles Tracy, then a young graduate of Yale College, afterwards a well-known lawyer in New York City, published a brief article on the subject, showing how the rotation of storms, which was at that time attracting much attention, should necessarily result from the rotation of the earth. This article was curiously overlooked; no reference was made to it till nearly forty years after its publication, and in the meantime the question had been much more fully investigated by others.

129. Ferrel's theory of the effects of the earth's rotation: 1856. Ferrel's studies of the effects of the earth's rotation on the circulation of the atmosphere were begun in 1856; they were more fully expanded in later years, and it is not too much to say that they have worked a revolution in the science of meteorology. Ferrel was at that time teaching school in Nashville, Tenn.; he was a self-taught mathematician of remarkable originality. Upon reading a statement of the erroneous theories of the winds then in vogue, he studied the matter out for himself and produced the first rational theory of the general circulation of the atmosphere. At that time the prevailing low pressure near the poles was coming into notice, particularly from the observations made in

the Antarctic voyages of Ross and Wilkes; but it received no proper explanation until solved by Ferrel, who showed conclusively that it must result from the establishment of an interchanging convectional circulation between the equator and the poles on a rotating earth.

It is impossible to present in brief outline and in non-mathematical form an adequate statement of Ferrel's theory; but the following paragraphs may serve to place its essential features before the student.

130. Motion on the rotating earth without friction. If a body be supposed to move without friction on the level surface of a rotating globe, a single impulse would give it a perpetual motion; but the motion could not be along a straight path. It would continually be deflected to one side of its momentary path, to the right in our hemisphere, to the left in the other, and with a force dependent on its velocity and on its latitude; but independent of its direction of motion. Its path would always be curved in a systematic manner; the curvature would be sharper for slow motions than for rapid motions, and sharper in high latitudes than near the equator. The following table will give some idea of the rate at which a moving body tends to turn from a straight line on a sphere rotating once in twenty-four hours; the first column being its velocity, the other columns giving the radius of curvature of the path in which it would move at several different latitudes.

RADIUS OF CURVATURE (IN MILES) FOR FRICTIONLESS MOTION ON THE EARTH'S SURFACE.

Latitude	0°	5°	10°	20°	30°	40°	50°	60°	70°	80°	90°
20 miles an hour	∞	880	442	224	153	119	100	88	82	78	77
10 miles an hour	∞	440	221	112	76	59	50	44	41	39	38
5 miles an hour	∞	220	110	56	38	30	25	22	20	19	19

A body once set in motion under these conditions would continue moving forever, always changing its direction but never changing its velocity. If it were given a velocity of 20 miles an hour in any direction at latitude 30°, it would describe a series of overlapping loops, gradually carrying it westward around the earth, but never passing outside of the parallels of 20° or 40°. If it were given a velocity of five or more miles an hour eastward at latitude 5°, it would describe a scalloped path, oscillating back and forth across the equator, but never escaping beyond latitude 5° in either hemisphere.

131. Movement of the air on gravitative gradients. The imaginary case of the preceding paragraph does not apply directly to the case of the winds, for they are not acted on by a single initial impulse, but by a continual gravitative acceleration, according to their gradient; and they are more or less

affected by friction and other resistances. The practical question then is: What will be the course of the general winds on a given gradient at a given latitude; it being understood that the conditions of steady motion have long ago been reached, as far as the average value of the gradients is concerned.

Consider for example the northeast trade wind of our hemisphere, AW, Fig. 31. It is urged by a small component of gravity, AC, to move towards the equator; but by reason of its velocity of about twenty miles an hour in latitude 20° N. it must experience a certain right-handed deflective force, represented by the arrow AD, at right angles to AW. The force, AC, expends one of its components, AF, in overcoming the deflective force (inertia), AD. The other component, AB, must just equal and oppose the sum of all the resistances, AE, that are excited by the velocity, AW; and in such a condition of equilibrium all the permanent winds of the world must blow. They always adjust their velocity and direction so as to bring about an equilibrium among the acting forces.

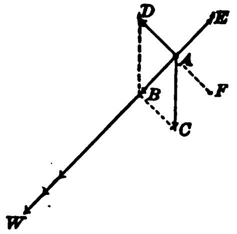


FIG. 31.

The prevailing westerlies of the far southern latitudes may receive a similar explanation; their austral deflection turning them to the left of their gradients.

Consider next the condition of the equatorial overflow that is inferred to run from the warm, expanded air above the equator towards the cold, compressed air of the south polar region. What course must it have in latitude 25° or 30° S. and at a great height above sea-level? An examination of the gradients of this part of the atmosphere, as illustrated in Fig. 26, shows that the accelerating force there must be much greater than that by which the trade winds are impelled. It may be represented by AC, Fig. 32. Moreover the resistance, AE, which the currents of the upper air encounter must be comparatively small. Now we must suppose that as the overflow current has certainly ages ago acquired the condition of steady motion, this small resistance is the equal and opposite of an equally small residual force, AB, left over from the gravitative acceleration, AC, after its greater part has been expended in overcoming the deflective force (inertia), AD. The part of gravitative force that acts effectively on a baric gradient is therefore even less than was indicated in Section 113. We may next proceed to determine the deflective force; and then

FIG. 32.

knowing the deflective force, the velocity and direction of the motion that arouses it may be found.

The deflective force may be found by completing the parallelogram, of which *AC* is a side and *AB* is a diagonal; the angle, *DAB*, being 90°. The deflective force, *AD*, is therefore almost equal and opposite to the gravitative force, *AC*. The overflow must then be represented by *AW*, showing it to have a high velocity and a direction but little south of east. In no other case can the deflective force have its appropriate direction and value with respect to the current that produces it. The greater part of the high level overflow from the equator must therefore run at a high velocity in a direction almost from west to east, but a little inclined toward the pole; in no other direction can the conditions of steady motion be reached in the presence of the small resistances of the upper air. All that is known of the high-level currents in middle latitudes from observations of clouds confirms this explanation in a striking manner.

132. Deflection of the winds from the gradients. One of the deficiencies in the convectional theory of the winds, pointed out in Section 125, is thus satisfactorily accounted for. Not only the trade winds and the prevailing westerlies follow the explanation of the preceding section, but all the smaller members of the general circulation also turn aside from their gradients, as seen in the spiral outflow from the South Atlantic area of tropical high pressure in January; or from that of the North Atlantic in July; or from the Asiatic area of continental high pressure in January; or as seen again in the spiral inflow towards the area of low pressure over the North Atlantic near Iceland in January; or over Asia in July; or over Australia in January. All these and many other examples to be met on subsequent pages are reconciled when the theory not only takes account of the interaction of insolation and gravity, but when it includes the effect of the earth's rotation as well.

Before advancing further in the discussion of the circumpolar whirl, an experimental review may be made of the points thus far learned.

133. Experimental illustration of the deflective effect of the earth's rotation. Stretch a smooth sheet of paper over a circular table, two or three feet in diameter, supported at the center on a vertical axis on which it may rotate in either direction. Lay a marble, dipped in ink, at the center of the stationary table. It remains at rest. This corresponds to the conditions of level isobaric surfaces, on which there could be no winds.

Set the marble rolling by a light blow; it will trace its path by dots of ink along a straight radial line; this corresponds roughly to the case of motion started by an initial impulse, on a smooth non-rotating earth. A body m...

under such conditions on the earth's surface would follow a straight path, always moving in the same great circle on which it started.

Rotate the table slowly from right to left, and set the marble in motion as before; it will describe a curved path, turning to the right of the radius on which its motion began. This corresponds to motion started in the northern hemisphere by an initial impulse on the level surface of a rotating earth. When the table is rotating at a given rate, a slow motion of the marble causes a sharp curvature in its path. When the marble is moving at a given velocity, a faster rotation of the table (corresponding to a higher latitude on the earth) causes a sharper curvature of the path. These results correspond essentially with those given in Section 130, although the experiment is imperfect from the effects of friction.

Tilt the table slightly on a hinge at the axis, so that it shall present an inclined surface, although it may still rotate on a vertical axis. Let the table stand still, and release the marble at the center. It will describe a straight path under the acceleration of the component of gravity which acts down the inclination of the surface. This corresponds to the gravitative acceleration of the wind on inclined isobaric surfaces on a non-rotating earth. The velocity attained will depend on the inclination of the table and on the resistances encountered. At a given inclination, the rougher the table, the slower the velocity when steady motion is gained. This corresponds to the conditions of steady motion in a simple convectional motion, as explained in Section 94. In ordinary experiments the table is too smooth and its radius too small for the attainment of steady motion.

The table being gently tilted, rotate it slowly from left to right; release the marble as before, and it will describe a curved path, turning to the left of the table gradient. This corresponds to the actual case of motion of the air on inclined isobaric surfaces in the southern hemisphere.

After the marble has rolled a little distance from the center, its path maintains an almost constant deflection from the gradient of the table. The larger and smoother the table, the better this may be seen. This corresponds to the conditions of steady motion under the action of gravitative acceleration and the deflecting force, as explained in Section 131. If the rotation of the table be slow, the final angle of deflection will be small, corresponding to the winds in low latitudes, where they do not turn far from the direction of the decrease of pressure. If the rate of rotation be faster, the angle of deflection becomes greater, corresponding to the strong departures of the wind from the gradient in high latitudes. If the table be smooth, the resistances to motion will be small, and the angle of deflection becomes large, as in the case of high-level atmospheric currents. If the table be rougher, a considerable value is required in the forward-acting resultant, and hence the deflection is comparatively small; this being the case of the lower winds, in contact

with the surface of the ocean, where they beat against the waves; and still more where they blow over the uneven surface of the land. This explains the prevailing difference between the direction of the surface wind and that of the lower clouds at a height of one or two thousand feet; the latter as a rule come, in our hemisphere, from a point somewhat to the right of the former, showing that their deflection from the gradient is somewhat stronger than that of the surface winds.

It is not at first apparent why the rotating table employed in these experiments may be compared with one or another part of the earth's surface. This may be made clear by the following illustration.

Several circular discs of paper, an inch or two in diameter and each marked with a strong diametral line, may be attached to a terrestrial globe in different latitudes. Watch the diametral line on one of the discs while the globe is slowly rotated; the line will be seen to change its direction; now pointing to one part of the room, now to another. In other words, the disc is rotating with respect to its center, and in the same direction as the globe rotates. A disc near the pole will rotate rapidly; a disc near the equator will turn its diameter more slowly from one direction to another; a disc on the equator has no motion of rotation with respect to its center; and at the equator there is no deflective force. These discs may represent the table of the preceding experiments. The marble at the center of the table or the air at any point on the earth experiences a deflection from its first path as it moves in any direction from the starting point; the deflective force varies with the velocity of motion and with the rate of the rotation of the surface with respect to its center; and the direction taken when steady motion is attained changes accordingly.

Hence, whenever a mass of air is impelled to move, as by the introduction of some change of temperature which produces a gradient in the previously level isobaric surfaces, it will soon turn from the gradient and adjust itself to an equilibrium under the action of gravitative acceleration and the deflective force. In motions between the equator and the poles, where the differences of temperature were long since introduced, and now vary only by moderate amounts above or below their mean value, the currents of the atmosphere must have long ago attained a condition of steady motion, hurrying a little when the differences of temperature increase in winter and steepen the gradients by a small amount, and falling to lower velocities when the gradients are weaker in summer. Inasmuch as the gradients and the deflective force vary from latitude to latitude, it follows that the velocity attained by the general circulation of the atmosphere must likewise vary between the equator and poles; but at every latitude a temporary equilibrium is taken, to be lost only as the wind moves into a new position, where the forces acting on it change their relative values.

134. Analogy with an eddy in water. An illustration of the atmospheric whirl mentioned in Sections 126 and 132 may be found in a basin of water discharging itself by a vent at the bottom. If the water stand still when the vent is opened, its currents will move in radially towards the center, and there descend without attaining any great velocity; but if a gentle rotary motion be given to the water before the vent is opened, the discharge will require a much longer time, and will be deflected so as to form a rapidly whirling central eddy or vortex of increasing velocity towards the center, where its centrifugal force may be so great as to open an empty core. The analogy of this case with that of the circumpolar whirl of our atmosphere is very imperfect; but it serves to emphasize the great value and strong effect of the centrifugal force that may be developed in such a vortex; and it is on such a centrifugal force that we are counting to reduce the expected high pressure at the pole into the actual low pressure.

If we imagine ourselves looking at the earth so that the south pole appears in the center, while the equator forms the marginal circumference, then the great equatorial overflow, rotating with the earth at the equator, may be likened to the case of the rotating body of water about to discharge itself at the center of the basin. Let us examine into the velocities and accompanying centrifugal forces that would be gained if there were no loss by friction.

Let the radius of the basin be ten inches; let the linear velocity of rotation at the margin of the basin be one foot a second. As the water is drawn in towards the center, its linear velocity of rotation will increase just as much as its radius is diminished: this is in accordance with a well-known mechanical principle, called the *conservation of areas;* because the area swept over in a given time by a radius from the center to any particle of water is constant. The centrifugal force developed by a rotating body varies with the square of the velocity of rotation divided by the radius of rotation. The following table exhibits the rapid increase of centrifugal force as the center of the vortex is approached.

CENTRIFUGAL FORCE IN A VORTEX.

RADIUS.	LINEAR VELOCITY	CENTRIF. FORCE.
10	1	$\frac{1}{10}$
5	2	$\frac{4}{5}$
1	10	100
$\frac{1}{10}$	100	100,000

It is manifest that even a less excessive centrifugal force close to the axis of the eddy may suffice to open an empty core, whose surface is everywhere at right angles to the resultant of the downward gravitative force and the outward centrifugal force.

135. Vorticular circulation of the atmosphere around the poles. We have now to inquire whether the low pressure of the atmosphere around the poles may result directly or indirectly from the cause by which the oblique course of the winds has been so successfully explained.

For this purpose the surface winds of the temperate latitudes and the high level currents above them, sidling swiftly along on their steep poleward gradients, must all be considered together. They combine to form a vast aërial vortex or eddy around the pole. In the northern hemisphere this great eddy is much interrupted by continental high pressure in winter or low pressure in summer, and by obstruction from mountain ranges, as well as by irregular disturbances of the general circulation in the form of storms, large and small. In the southern hemisphere the circumpolar eddy is much more symmetrically developed. What effect will be produced on the pressure of the atmosphere in high latitudes by the whirling of the great body of air around the pole as a center?

136. Cause of low pressure around the poles. If the explanation of Section 134 be now applied to the atmosphere, with the supposition that there is no loss of velocity by friction or other resistances, it is clear that an excessive velocity and a still more excessive centrifugal force would be developed in the circumpolar vortices. It should be noticed that the eastward motion of 1,000 miles an hour that the air has over the equator is increased as the overflow approaches the pole; at latitude 60°, where the distance from the axis is half what it was at the equator, the eastward velocity has doubled; that is, it has become 2,000 miles an hour, or 1,500 miles faster eastward than the earth's surface at that latitude. Forty miles from the pole, it would be 100,000 miles an hour; and so tremendous a velocity on so short a radius would suffice to hold the air away from a closer approach to the pole, if it could, indeed, approach so close as this; at any less distance there would be a vacuum.

But the action of friction and other resistances cannot be neglected. The presence of almost as great an atmospheric pressure in the polar regions as at the equator assures us that the imaginary case of no friction is far from the actual case. Although the resistances suffered by the upper air currents cannot be great, they successfully prevent the realization of the enormous circumpolar velocities that would result in the case of no friction and no intermingling of currents. The reason for this is seen in considering again the conditions of steady motion represented in Fig. 32. The overflow takes so nearly an eastward direction that it must travel over a very long distance in advancing from the equator to the pole; and in all this distance, the theoretical increase of velocity with decrease of radius is continually defeated by the action of the small resistances.

Excessive velocities cannot be reached; but from observations on high-level clouds, it is known that the currents five or six miles above sea-level frequently move to the eastward at a rate of a hundred or more miles an hour; it is very probable that much greater velocities would be encountered at heights of ten or fifteen miles. Although the mass of the atmosphere may be divided into a lower and an upper half at a height of about three miles, it is calculated that half of the capacity of all the atmospheric currents for doing work — a function of their volume, density and velocity — is not measured until a height of about eight miles is reached; the great velocities of the still higher but much smaller mass of air giving it a capacity for work equal to that of the slower moving but much greater mass below this height. When it is remembered that the eastward velocity is in excess of the already rapid eastward movement of the earth's surface, it will be seen that the deflection towards the equator arising from the abnormal centrifugal force thus developed may be fairly accepted as the cause of the unexpected low pressure around the poles.

On a rotating earth, the convectional circulation between the equator and the poles cannot follow the meridians. It must suffer deflection into oblique paths and thus develop a vorticular whirl around the poles.

The high pressure that should result from the low polar temperatures is therefore reversed into low pressure by the excessive equatorward centrifugal force of the great circumpolar whirl; and the air thus held away from the polar regions is seen in the tropical belts of high pressure. Thus the second deficiency of the convectional theory of atmospheric circulation is accounted for as satisfactorily as the first. The theory may be regarded as firmly established.

The credit of first explaining the greater movements of the atmosphere and the general distribution of pressure in accordance with just physical principles belongs to Ferrel. Following the results first reached by him, others have since then confirmed his chief conclusions and extended their researches to a fuller statement of meteorological problems. Chief among these later investigators may be mentioned Oberbeck, whose studies have been concerned with the general movements of the atmosphere; and Helmholtz, who has shown how the general movements may provoke subordinate movements, in the manner briefly referred to in sections 203 and 237.

CHAPTER VII.

A GENERAL CLASSIFICATION OF THE WINDS.

137. Basis of classification. Having now come to a general understanding of the convectional circulation of the atmosphere on the rotating earth, a systematic account of the different members of the circulation may be attempted. A classification proposed by Dové, an eminent German meteorologist of the first half of this century, divided the prevailing winds of the world into three classes: the permanent winds, of which the trades are the chief members; the periodical winds, of which the monsoons of India are the type; and the variable winds, comprising the irregular but prevailing westerly winds of middle latitudes. While this classification has been generally adopted, it does not satisfy the present demands of our science, being both arbitrary and incomplete, and failing to recognize the natural relation that exists among the various movements of the atmosphere. A classification according to cause is here presented in preference.[1] It is true that we ordinarily take no account of the difference of cause between a gentle breeze of mild temperature and a violent winter gale, preferring to classify them according to their direction, velocity or temperature; but in seeking for an explanation of the phenomena of the atmosphere, it is advisable to consider the various causes of motion separately. The classification of winds presented in the following table therefore is arranged, first, according to the source of the energy on which they depend, and second, on the manner or period of its application.

138. Classification of winds according to cause.

Source of Energy.	Application.	Period.	Name of Wind.
Solar heat . . .	Equator and poles . .	Permanent	Planetary.
" " . . .	Heat equator and poles .	Annual . .	Terrestrial.
" " . . .	Continents and oceans .	Annual . .	Continental.
" " . . .	Land and water . . .	Diurnal . .	Land and sea breezes.
" " . . .	Mountains and valleys .	Diurnal . .	Mountain and valley breezes.
" " . . .	Local, or indirect . . .	Irregular .	Cyclonic and other storms.
" " . . .	Light and shadow . . .	Irregular .	Eclipse winds.
" " . . .	Indirect	Accidental .	Landslide and avalanche blasts.
Lunar attraction .	Through the tides . . .	Twice in a lunar day .	Tidal breezes.
Telluric heat . .	Volcanic eruptions . .	Irregular .	Volcanic storms.

[1] Following closely the classification published by the author in the American Meteorological Journal for March, 1888.

139. Tidal breezes. The insignificance of all winds other than those of direct solar origin will be perceived by glancing at those here referred to lunar, telluric and indirect solar causes. Under lunar winds may be placed those light movements of the air which are thought to correspond with and appear to be determined by the tides, where the rise and fall of the sea-surface, as in estuaries, is of considerable amount. The air is raised and pushed away by the rising water; and when the water sinks, the air is drawn down after it. Near the borders of an estuary with strong tides, it is said that winds of this class are perceptible. They are said to occur in the Gulf of St. Lawrence, but they are of subordinate quality and are generally overcome by stronger winds of other kinds. The strong tides of the Bay of Fundy should produce winds of this class, if they occur anywhere: they would be detected by hourly records of the wind direction at several stations around the Bay, tabulated according to the lunar day, instead of the solar. It is believed that land and sea breezes are intensified at certain tropical stations, when they move with the falling and rising tide. It may be here mentioned that the most careful analyses of wind records at inland stations, or over the world in general, have failed to detect anything more than the faintest and most questionable control of the winds by the moon, except in the indirect manner above indicated. The more closely the subject is investigated, the less reason there appears to be in the popular belief that the moon exercises any significant control over the weather.

140. Volcanic and accidental winds. Winds are occasionally associated with volcanic outbursts; either from the explosive action of the eruption or from the convectional motion of the air over incandescent lavas. These are the only winds of purely telluric origin, and need only to be mentioned to show their rarity.

Destructive blasts of air are sometimes brushed forward by landslides or avalanches; they may be of sufficient violence to overturn houses and trees many hundred feet in advance of the slide. Although of telluric origin at first sight, landslide blasts should be regarded as of indirect solar origin, inasmuch as landslides in all cases depend on the erosive action of rain and streams, and these depend on sunshine. Avalanche blasts are more apparently of indirect solar origin.

Certain observers have reported a light wind moving from the space traversed by the passing shadow of the moon during a solar eclipse, as if the air under the shadow became somewhat cooled by radiation, and thus developed a faint convectional descent and outflow. In the event of a total solar eclipse occurring in a populous country and at a time of quiet weather, hourly observations of the wind might be undertaken by numerous observers to determine the extent of this peculiar member of the family of winds.

All the lunar, volcanic, accidental and eclipse winds together are of the most trifling value compared with the vast systems of prevailing winds embracing the whole earth in their circuits; of continental winds, sweeping in to and out from the center of even the largest land areas; or of stormy winds, whirling in cyclonic eddies a thousand miles in diameter, travelling for a week or a fortnight, and crossing lands and oceans on their way. These are all winds of solar origin.

141. Planetary winds. All rotating planets that have an atmosphere and are warmed around the equator by a sun must possess more or less perfectly an oblique circulation between the equator and poles, of a kind already outlined in previous sections. The essential features of such a circulation may now be concisely stated.

Supposing the surface of the planet to be smooth, and the contrast of temperature between equator and poles to be strong, there will be a belt of low pressure around the equator, tropical belts of high pressure at some intermediate latitude, and caps of low pressure over the poles. The arrangement of the isobaric surfaces of the atmosphere would correspond to that for the southern hemisphere in Fig. 26. The overflow from the warm equator would soon turn forward in the direction of the planet's rotation ("eastward") and thus develop a rapid whirl around either cold pole. On account of the convergence of the meridians towards the poles, much of the air that departed from the equator would return in an under-current at various intermediate latitudes; only the smaller share would complete the entire circuit.[1] As the branches from the overflow descend to lower levels to begin their return course, they encounter gradients still directed towards the poles, but less steep than those aloft. The strong equatorward deflective force gained by the currents on the steeper upper gradients serves to carry them against the slope of the weaker lower gradients; thus they return obliquely towards the equator. But if the currents descend close to sea-level, their velocity is so much reduced by friction that they obey the gradients and run obliquely towards the poles, forming the prevailing westerly winds of middle and higher latitudes.[2]

[1] The tropical belts of high pressure are sometimes explained as a result of the crowding of the equatorial overflow as it advances along the converging meridians. This is incorrect. If the convergence of the meridians determined an increase of pressure, the pressure should be highest at the poles where the meridians converge most rapidly. The convergence of the meridians has no significant share in the increase of pressure towards the pole. The tropical belts are due essentially to the high temperature on the equatorial side and the deflective force of the circumpolar whirl on the polar side.

[2] An illustration from another point of view may make this matter clearer. In saying that the return currents go back to the equator against the gradients, it might be asked: what is the standard of "level" from which the slope of an isobaric surface is determined? The standard to which we are accustomed is the surface of the sea; but it must be remembered

142. The members of the planetary system of winds. This theory of the planetary circulation therefore demands the existence of an overflow, approaching the pole in a spiral course; an intermediate return current, supplied at all latitudes by branch currents descending in short circuits from the lofty overflow and receding from the pole in a spiral course, but still moving

that this surface is thirteen miles further from the center of the earth at the equator than at the poles, and that it might therefore be said in a certain sense to ascend from poles to equator. The reason that we call it level is that we are guided in the determination of a level surface only by our knowledge of the direction of gravity, to which a level surface must be perpendicular. Gravity, however, is not directed towards the center of the earth, as has been explained in Section 13. In the youth of the world, when its rotation was presumably faster than now, a different idea of "level" must have obtained.

For the same reason, the idea of "level" that the Circumpolar Eddy possesses cannot agree with ours; for its period of rotation around the earth's axis is less than twenty-four hours. From the Eddy's point of view, the surface of the ocean, which we call level, must slant down towards the equator. Indeed, the isobaric surfaces at the level of the return current, which we say slant to the pole, must seem to the Eddy to slant faintly towards the equator; and with this understanding of gradients, it is natural that the return current should follow what it regards as their direction of descent.

Figure 33, corresponding to Fig. 26 of Section 111, may make this still plainer. The meridional components of the winds are shown by dotted lines for a quadrant of the section. To us, who live on the surface of the rotating earth, the elliptical meridian, NQS, appears "level." If the earth did not rotate, this meridian would be called "up hill" toward the equator, because it rises in that direction above the circular meridian, nqs, which would then be called level. The rapidly whirling equatorial overflow regards all the isobaric surfaces above the line, ABC, as slanting towards the poles, for in spite of its excessive equatorward deflection, the upper surfaces descend so rapidly towards the pole that their descent must be recognized. But when the branches of the lofty overflow descend to enter the return current, beneath ABC, they find isobaric surfaces that are not so steep towards the poles; and these they mistake (as we should say) for gradients directed towards the equator. The line, ABC, therefore represents the "neutral plane" of the planetary circulation. Finally, the surface winds, having only a moderate velocity eastward in excess of the earth, see the lower gradients as we do, and sidle along them towards the pole.

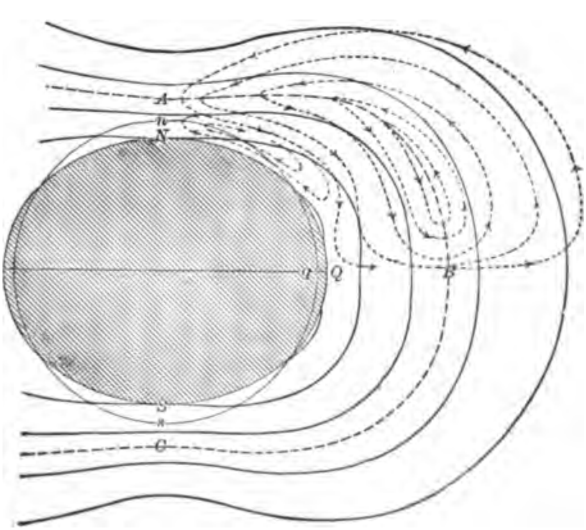


FIG. 33.

forward in the direction of the planet's rotation until it passes the axis of the tropical high-pressure belt, where it moves obliquely backward as the trade wind; and a lower current, constituting the prevailing westerly winds of the middle and higher latitudes, approaching the pole in a spiral course, like the overflow aloft. The intermediate and upper members are not yet clearly distinguished by direct observation, for the actual circulation of the earth's atmosphere is much disturbed by continental obstruction and by stormy overturnings. Indeed, there is reason to believe that the confusion of currents thus introduced constitutes the greater part of the resistances encountered by the lofty currents of the equatorial overflow.

Calms should occur at all places of no gradient; that is, along the axes of the equatorial and tropical belts and close about the two poles. The two former are well confirmed by observation, and record of the two latter may be expected when the poles are explored.

Along the equator, above the surface calms, the trade winds converge and move from east to west, ascending obliquely as they go, and gradually turning north or south when they mount to an altitude at which the poleward gradients appear. The overflow therefore begins with a westward component not before mentioned; but this is soon lost under the action of the deflecting force proper to the hemisphere that the air enters; and the current swings around toward the east.

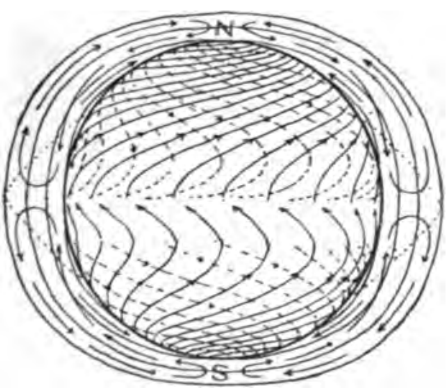


Fig. 34.

An ideal planetary circulation is represented in Fig. 34. The upper currents are drawn in full lines in the northern hemisphere, with the intermediate return currents in dotted lines. The return currents are drawn in full lines in the southern hemisphere, with the surface winds in turn dotted beneath them.

An essential characteristic of the oblique planetary circulation is the retardation of the atmospheric interchange between the equator and the poles, as compared with the rate it would attain on a non-rotating planet on which the winds would follow the meridians. For while the circumpolar winds on a rotating globe reach much higher velocities than would be gained by the meridional winds of a stationary globe, the oblique course of the circumpolar winds reduces their meridional components of motion below the values they would have in the direct north and south circulation of a stationary globe. The contrast of polar and equatorial temperatures is therefore greater on a rotating than on a non-rotating planet under similar supplies of insolation.

The members of the circulation of our atmosphere, which may be taken as illustrating the planetary system of winds, may now be described. It is natural that they are found in best development over the oceans; while the continental areas cause interruptions which will be considered further on.

143. The trade winds blow between the tropical belts of high pressure towards the equatorial belt of low pressure; from the northeast in this hemisphere, and from the southeast in the other. They are best observed on the oceans, where they hold their courses steadily over great areas, the most regular winds of the world, with a brisk velocity on gentle gradients. They occupy nearly half the earth's surface, and thus add to the uniformity of the great torrid zone, already signalized by its faint contrasts of temperature and its small change of seasons. Their name comes from their steadiness, and not, as the dictionaries sometimes say, from their benefit to commerce. Their course is so steady that it has given name to the Windward and Leeward Islands of the Lesser Antilles. These great streams of air average two miles or more in depth; only the higher mountains of their latitudes rise above them into the oblique westerly currents or anti-trades of still higher levels. They are seldom invaded by storms; they bear only small clouds by day and at night they are nearly cloudless; their mass warms throughout and expands to greater volume as they draw nearer the equator; and they gather a great amount of vapor from the ocean surface as they brush the waters along in their course. Yet while the winds at sea may blow for days or even for weeks with slight variation in direction or strength, their uniformity as displayed on our planet should not be exaggerated. There are times when the trade winds weaken or shift; there are regions where their steady course is deformed, most notably about the larger island groups of the Pacific; the Fiji and Samoa or Navigators Islands. At certain seasons in the several oceans they are invaded by revolving storms or cyclones of terrific energy (Sect. 217). Near the coasts the trades are interrupted by the daily breathing of the land and sea breezes; they are greatly deflected by mountain ranges and continental barriers; and over the torrid lands they may be entirely broken up for part of the year by the strong seasonal changes of temperature.

144. The doldrums or equatorial calms lie between the steady trades along the barometric equator of no gradients: a belt of light and variable winds and frequent calms, with cloudy, rainy sky, accompanied by thunder-storms and squalls. The air of the moist trade winds, flowing in obliquely from either side, here loiters about, and were it not for the shouldering off of the lofty overflow by the expansion of the lower air, the equatorial low-pressure belt would soon be filled up. Sailors find that the doldrums, with their calm, sultry air, their light and baffling breezes and frequent rains, stand in dis-

agreeable contrast with the refreshing air of the trades. In the doldrums the ocean may be glassy smooth, reflecting the gray or leaden color of the clouds; the sails hang lazily from the spars, waiting for a chance breeze, flapping only as the vessel rolls in the long swells that swing across the sea from stormier latitudes. In the trades the sea is roughened under the brisk wind; its color is clear dark blue; ships sweep swiftly across it, holding close to their course, every sail full and drawing, and the vessel steadily canted to leeward day and night.

145. Horse latitudes. The vague tropical belts of high pressure on the outer margins of the trades are characterized by light, variable winds and occasional calms, known at sea as the *horse latitudes* or tropical calms; but unlike the doldrums the weather here is comparatively clear and fresh. The meaning of these marked contrasts, to be given in the chapter on rain, will be found to afford correlations of much value in testing the general theory of atmospheric circulation.

146. Prevailing westerly winds. Outside of the tropical belts of high pressure, all across the temperate zone and even into the frigid regions, we find the *prevailing westerly winds;* the surface members blowing west-southwest in this hemisphere, west-northwest in the other. As the upper and lower currents are here of nearly the same direction, the westerlies are of greater depth than the trades; they prevail to the summits of the highest mountains and even to the level of the loftiest clouds. They are frequently interrupted by storms, rare in the trade-winds of the torrid zone; indeed, the winter of our temperate zone on land is characterized by so continuous a succession of shifting winds that the prevailing direction of movement is hardly apparent until a careful record is examined. The interference of mountains and continents with the westerlies of the northern hemisphere is even greater than with the trades; and for this reason, the southern hemisphere, where the temperate latitudes are nearly all of ocean surface, offers a much better field for the study of the westerlies, as members of the planetary circulation, than is found nearer home.

Between latitudes 40° and 60° south, the "brave west winds" blow almost continuously from some westerly point; shifting somewhat when compounded with passing cyclonic whirls, especially in the higher latitudes, and often holding the strength of a gale for days together, particularly in winter; if reversed to an easterly direction, the wind remains in that quarter for but a brief time and is seldom violent. For this reason, beating around Cape Horn to the westward is dreaded by sailors; the wind there is boisterous and stormy, the air cold and wet, the sea nearly always rough. Sailing vessels from England to the Australian colonies therefore take an outward course by Cape

of Good Hope, but return across the South Pacific, passing Cape Horn to the eastward, thus carrying fair winds nearly all around the world.

In passing to still higher latitudes, about 60° in the northern hemisphere and somewhat further from the equator in the southern, winds from a polar source become more common; very little is known of them, and as they seem to be due to disturbance in the normal planetary winds caused by the lands, they will be briefly referred to in a later section.

147. The upper currents blow prevailingly from the west and with high velocity in nearly all latitudes. Observations of the loftiest clouds disclose their rapid movement and almost constant drift from some westerly point. Temporary departures from this direction in temperate latitudes are always associated with some passing cyclonic disturbance. On Pikes Peak, over 70 per cent. of the winds recorded in ten years of observation came from some westerly point, and 36 per cent. from the southwest. Even in the trade-wind belt, the winds on lofty peaks, the carriage of volcanic smoke and ashes, and the drift of the upper clouds indicate as steady a westerly wind or anti-trade aloft as easterly below; the upper wind turning obliquely from the equator while the lower wind approaches it; and the two together undoubtedly forming compensating members of the terrestrial circulation. The anti-trades are felt on the summit of Teneriffe and on the volcanoes of the Hawaiian Islands, while clouds are seen floating in the trade-winds below.

But close to the equator, the upper currents are from the east. This is known from occasional observations on cirrus clouds, and better still from a peculiar effect of the great volcanic outburst of Krakatoa in 1883 already referred to. The dust and vapor blown out by the volcano caused remarkable displays of sunset and sunrise colors (Section 71); these were noticed at places progressively further and further west around the equator, encircling the world in fifteen days; thus determining a westward transportation of the dust at the rate of seventy miles an hour. After encircling the earth, similar brilliant sunsets were observed progressively northward and southward from the equator, as if the dust were then gradually and broadly distributed by the poleward overflow of the upper atmosphere.

148. Winds on other planets. The telescope reveals markings on some of the planets of our system that are interpreted as cloud belts, arranged by their winds. The great planet Jupiter has a well-marked system of belts; their distinctness is doubtless in good part due to his rapid rotation in nine hours and fifty-six minutes, which would develop a strong deflecting force and cause a distinct "flattening" of his winds. The axis of Jupiter is so nearly at right angles to the plane of his orbit that his wind system can have little annual variation. Saturn, whose day is almost as short as that of Jupiter, also has

atmospheric belts of lighter and darker color, indicating an atmospheric circulation; his axis being inclined almost 27 degrees to his orbit, and his year being long, his wind system must shift north and south like ours. A planet like Uranus, whose axis is thought to be nearly coincident with the plane of his orbit, would, if the heat of the sun were sufficient at so great a distance, have an extraordinarily well-developed migration in his wind system: during its long spring and autumn, the circulation would be like that of Jupiter; but in its equally long winter and summer, the pole of the sunny hemisphere, with continuous sunshine for many of our years, would become the center of a great cyclonic whirl, with an ascending instead of a descending component of motion. Planetary winds thus appear to be of different kinds. We shall not fully appreciate the special features of the winds of our planet until the peculiarities by which they are distinguished from the Jovian and Uranian winds are clearly perceived.

149. Terrestrial winds. The general scheme of planetary winds may now be more closely adapted to the earth by considering, first, the effect of the inclination of its axis to the plane of its orbit; second, the disturbing effect of its continents and mountain ranges. This section will include only the first of these effects; and the planetary winds thus modified will be called the *terrestrial winds.*

The inclination of the earth's axis to the plane of its orbit causes an annual migration of the heat equator and an annual variation in the poleward temperature gradients, as has been explained fully in a former chapter. In consequence of this migration of the heat equator, the barometric equator, or axis of low equatorial pressure, must also migrate with its belt of light winds and calms, and when its extreme northern or southern position is assumed, the winds of the winter hemisphere must extend across the geographic equator for a little distance into the summer hemisphere. The effect of this on the course of the trade winds is remarkable. Consider the time of the late northern summer, when the barometric equator on the Pacific lies between 7° and 10° N. latitude. The southeast trade wind of the southern hemisphere is then led by continuous northward gradients across the equator; but on entering our hemisphere, and finding itself under the control of a right-handed deflecting force, it swings around and becomes a southwest wind, occupying a latitude belt that was traversed by the normal northeast trade wind six months before. Although not conspicuous, on account of the many irregularities to which the course of the winds is subject, the special charts of the Pacific clearly recognize the existence of this interesting feature of the terrestrial circulation. The opposed winds of this belt may be called the terrestrial monsoons of the Pacific; monsoon being a general term now applied to winds whose direction is reversed once a year. The equatorial counter-current of

the Pacific waters appears to be chiefly due to the action of this special member of the terrestrial circulation. A belt of similar terrestrial monsoons occurs south of the equator in the Indian ocean. When the terrestrial monsoons are aided by continental influences, alternating winds of much greater extension and distinctness are developed, as those of the North Indian ocean and the Chinese seas (Sect. 153).

The annual change in the value of the poleward temperature gradient causes a corresponding variation in the barometric gradient; and although this is of small amount, it has been recognized as a characteristic of the distribution of pressure on the isobaric charts (Sect. 114). In consequence of this, there is a distinct annual variation in the velocity of the prevailing westerly surface winds of middle and higher latitudes, and of the higher currents also, as determined at mountain observatories and by observations of clouds. Extended cloud observations at Blue Hill, Mass., indicate that the entire atmosphere, from the lowest to the highest cloud level, moves almost twice as fast in winter as in summer; the mean velocity of the highest clouds in winter being over 112 miles an hour, and the highest velocity determined being 230 miles an hour. The velocities determined by observations in Europe are not so great, but they show a similar annual variation. With increase of velocity in the winds, the equatorward deflecting force is increased; thus the air is held more effectively from the polar regions, and the tropical belt of high pressure in the winter hemisphere is driven further towards the equator after the retreating doldrums. When the winds slacken, the deflecting force is relaxed and the high-pressure belt moves towards the pole. The "horse latitudes" therefore shift north and south in an annual period sympathetically with the equatorial calms or doldrums. This will be found to exert a marked control on the rainy seasons of certain regions (Sect. 302).

150. Annual migration of the wind system. The whole system of surface winds shifts north and south after the sun, moving northward till August or September and southward till February or March, as is shown in the following table of limiting latitudes of the several members on the oceans.

	Atlantic Ocean.		Pacific Ocean.	
	March.	September.	March.	September.
NE. Trades . . .	26° N. — 3° N.	35° N. — 11° N.	25° N. — 5° N.	30° N. — 10° N.
Doldrums	3° N. — 0°	11° N. — 3° N.	5° N. — 3° N.	10° N. — 7° N.
SE. Trades . . .	0° N. — 25° S.	3° N. — 25° S.	3° N. — 28° S.	7° N. — 20° S.

The shifting is much less than the migration of the sun from the equator; and the maximum migration of the wind system, north and south, occurs one or two months after the solstices; an example of the belated occurrence of an effect after its cause.

151. Sub-equatorial and sub-tropical wind belts. In consequence of the shifting of the members of the planetary system that is seen in the terrestrial winds, it is advisable to introduce special names for those belts over which the equatorial and tropical calm belts annually migrate. The first may be called the sub-equatorial belt; as has already been stated, and as is now illustrated in Fig. 35, its northern half has alternate northeast and southwest monsoon winds; its southern half, southeast and northwest monsoons in the winter and summer of the respective hemispheres. The other belts are called the northern and southern sub-tropical belts, where the steady trades and the stormy westerlies alternately hold possession, summer and winter. It must be carefully noted that these smaller features of the terrestrial circulation are by no means regularly and symmetrically developed on the actual earth, however well they might be formed on an earth all covered by an ocean. Their boundaries are not on well-defined latitude lines, as in the diagram, but are often rendered vague and irregular by other than terrestrial causes; yet they are distinct in certain regions, and they will have repeated mention in later sections concerning storms, rainfall and climate.

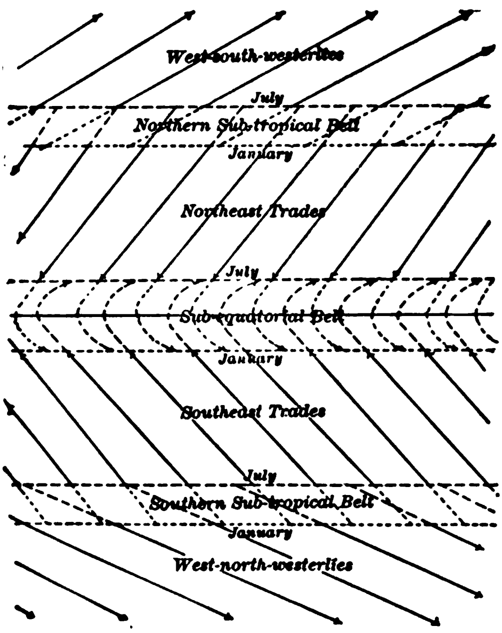


FIG. 35.

152. Continental winds. The interruptions in the terrestrial winds due to the action of the continents are of two kinds. One arises from the contrasts of temperature on land and water and acts in opposite directions, winter and summer: it would appear in full development, even if the lands were perfectly level and elevated above the sea only enough to prevent their submergence. The other depends on the obstruction offered to the terrestrial winds by the inequalities of the land, notably by the plateaus and mountain ranges: it always acts in the same way, and is analogous to the effect produced bv the continents on the currents of the ocean: it would appear, even if there were no differences of temperature between land and water.

The isothermal charts for January and July have already shown that the lands of the temperate zone are alternately warmer and colder than the adjacent oceans. They must therefore cause seasonal changes from low to high pressure,[1] and the terrestrial winds must be more or less affected by these changes, as has already been briefly referred to in considering the points of correspondence between theory and observation. The inflow towards the warm lands of summer and the outflow from the cold lands of winter will be appropriately deflected to the right or left according to the hemisphere. The overgrown continent of Asia presents the most striking illustration of winds of this class; the pressure over a large part of Central Asia, when reduced to sea-level, varies by eight-tenths of an inch from January to July, and the general course of the planetary winds is therefore greatly modified, as may be seen on Charts V and VI. In our western interior the annual variation is about four-tenths of an inch: in Australia it is about three-tenths of an inch. With these changes of pressure there are sometimes complete reversals in the direction of the winds, which are then called *continental monsoons.*

A small illustration of continental winds is found in the peninsula of Spain. The seasonal variations of temperature on the peninsula have already been shown in Figs. 14 and 15, Section 85. The effect of these changes of

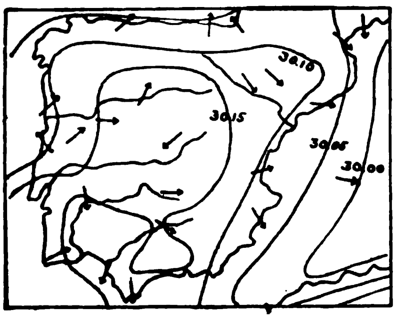


FIG. 36 (January).

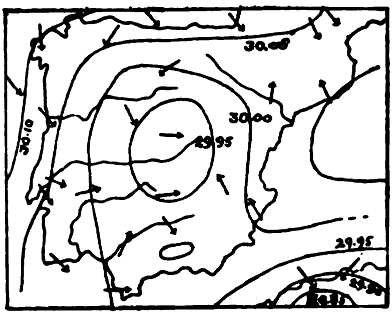


FIG. 37 (July).

temperature on the distribution of pressure and on the associated course of the winds is given in Figs. 36 and 37. The first shows January, with an interior area of high pressure and obliquely outflowing winds; the second, July, with an interior area of low pressure and obliquely inflowing winds.

153. The monsoons of Asia. In the winter season, when the equatorial belt of low pressure lies eight or ten degrees south of the geographical equator in the Indian Ocean, the abnormally low temperatures prevailing over Asia increase the extent and value of the baric gradients on which the trade winds

[1] See foot-note to Sect. 114.

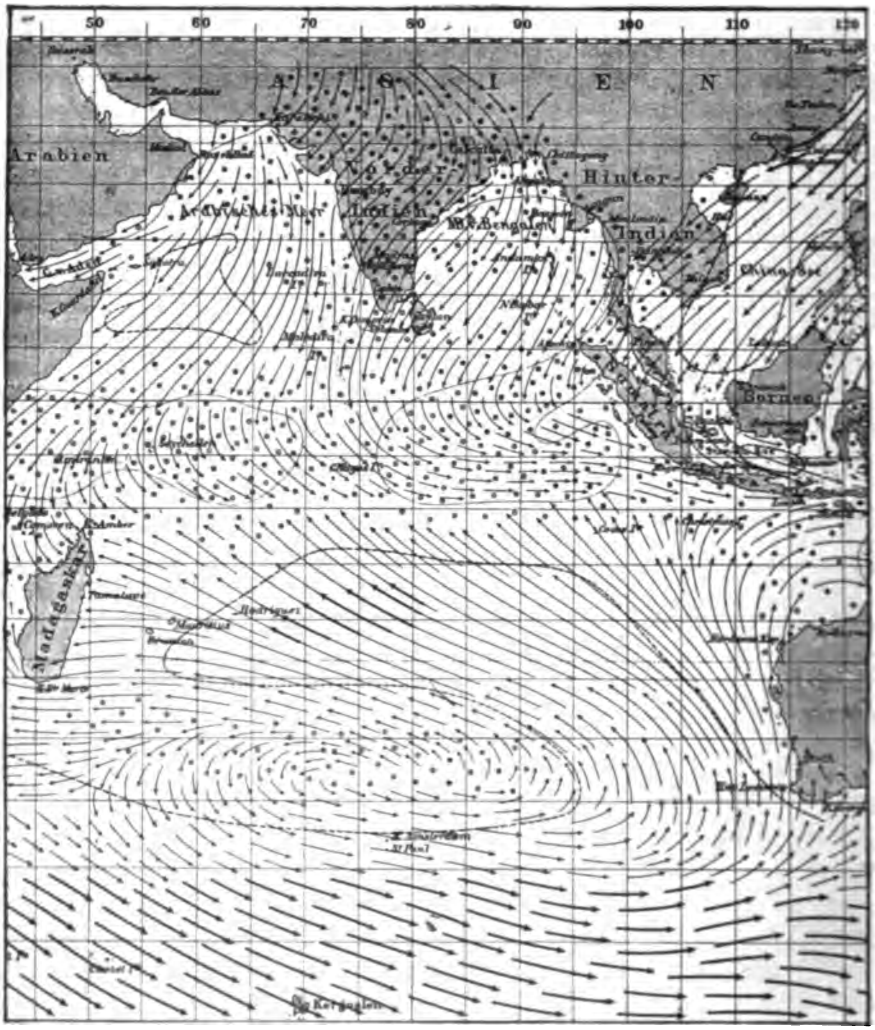


FIG. 38. — GENERAL WINDS OF THE INDIAN OCEAN FOR JANUARY AND FEBRUARY.

(From the Atlas of the Indian Ocean of the German Naval Observatory.)

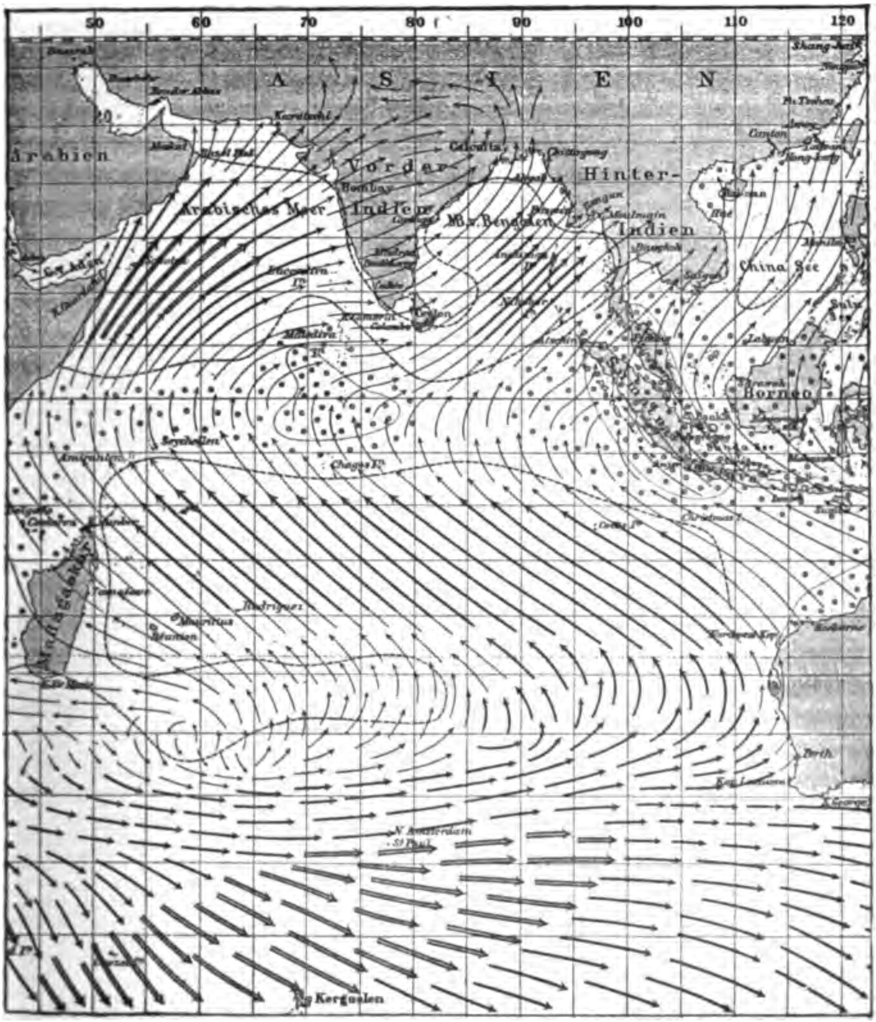


Fig. 39. — General Winds of the Indian Ocean for July and August.

(From the Atlas of the Indian Ocean of the German Naval Observatory.)

of our hemisphere blow; and all the southern and eastern coasts of that vast continent are swept over by an outflowing wind, generally from the north or northwest on the coast of China and from the northeast over India, but having a direction locally much influenced by the form of the land, and hence better named the winter monsoon than by a name indicative of its direction. As drawn in Fig. 38,[1] the winter monsoon of northern India flows from the northwest in the plain of the Ganges. Over the land, this wind is generally weak, cool, and dry; in China, when reinforced by stormy disturbances, it may become strong and cold. Over the sea, it blows briskly and takes the normal direction of the northeast trades, crossing the equator on its way to the belt of calms; but shortly after entering the southern hemisphere, it appropriately turns to the left of the gradients and blows as a northwest wind, a true terrestrial monsoon; not so steadily here as farther north, and yet appearing distinctly enough in charts of the average wind direction of these special latitudes.

In the summer season, Asia is the seat of unduly high temperatures, and by the aid of its numerous lofty mountains and plateaus a great depth of atmosphere is warmed abnormally. The high pressure of winter is then reversed into a low pressure, and the winds blow inward from all sides, even from the South Indian and the Arctic oceans. Over India, the general direction of this monsoon is from the southwest; but it is turned to the southeast on the plain of the Ganges. On the coast of China, it comes from the south or southeast. It is a warm, sultry, moist and rainy wind, of decidedly greater velocity than the winter monsoon.

The monsoons of India are the most famous winds of their class. Their name is derived from an Arabic word, meaning season. The belt of low pressure that lay to the south of the equator in our winter migrates gradually northward in spring, and is finally replaced by the formation of an area of low pressure over the warm desert plains of India and Persia, even as early as May. A northward gradient then leads the southeast trades of the South Indian ocean across the equator, and on entering our hemisphere they swing around and blow from the southwest, as shown in Fig. 39; thus presenting one of the most interesting phenomena in the circulation of the atmosphere. The northern Indian ocean and the adjacent seas are thus alternately swept over by winds of northeast and southwest directions, and on land these winds dominate the change of the seasons. The true terrestrial monsoons are seen on a belt of the Indian ocean close south of the equator; being occupied by the normal

[1] Figs. 38, 39, 40, and 41 are copied from charts prepared by Dr. W. Köppen for the German Naval Observatory (*Deutsche Seewarte*) at Hamburg. The first two are taken from the Atlas of the Indian Ocean; the second two, from the Sailing Handbook of the Atlantic Ocean. Long wind arrows denote steady winds; shorter arrows, variable winds. Heavy wind arrows denote gales or strong winds; light arrows, moderate winds; small circles indicate calms. A fuller account of these charts is given in the American Meteorological Journal, vols. IX and X.

southeast trade in our summer, and by the deflected extension of the northeast trade which turns northwest after it crosses the line in our winter. The monsoons north of the equator are a product of terrestrial and continental winds combined.

154. Monsoons of Australia and elsewhere. Australia, alternately too warm and too cold for its latitude, and hence with pressures alternately lower and higher than those of the surrounding seas, possesses winds blowing spirally inwards in January and outwards in July. They are not, however, in all parts of this land reversed directly enough to deserve the name of monsoons. In January the equatorial belt of low pressure becomes confluent with the local area of low pressure over Australia, and the outflowing monsoon of the Chinese coasts crosses the equator and reaches Australia as a northwest monsoon; thus repeating the illustration already afforded by India of a continental deformation of the terrestrial wind system.

In the Atlantic ocean, north of the equator and adjacent to Africa, there is a small area alternately swept over by the northerly trades of winter and a deflected extension of the southerly trades in summer, thus producing a distinct monsoon-like reversal in the direction of the winds with the seasons. The same change appears to occur in equatorial Africa, as the belt of calms shifts north and south after the sun. As the wind there blows over the land and near the equator, its deflection from the gradients is small, so that in successive halves of the year it appears as a nearly reversed north and south wind, and hence of monsoon quality. In equatorial South America this change does not appear so distinctly; the winds of the plains of the Amazon coming chiefly from the east, probably by reason of the heavy rainfall that occurs along the eastern base of the Andes, towards which the air is drawn.

The winds on the Texas coast of the Gulf of Mexico manifest a distinct monsoon tendency, blowing more frequently from the south in the summer and from the north in the winter; but as their direction is complicated by the action of passing cyclones, they are not steady and they do not exhibit the strong and persistent seasonal contrasts found in the classic monsoons of India. Of all these examples, that of Asia and in particular of Southern Asia and the adjacent Indian ocean is the most instructive. The extension of the southeast trade-wind of the Indian ocean in the northern summer as a southwest monsoon in the area normally belonging to the northeast trade, and the reverse condition in southern summer, are the most remarkable occurrences of the kind in the world.

✓ **155. The monsoon influence on the terrestrial winds.** The cases of Asia and Australia already given illustrate the considerable deformation of the normal system of terrestrial winds by continental influences; the migration of

certain members of the terrestrial system being greatly increased by the action of the land in causing a strong migration of the heat equator. In most parts of the world, however, the effects of the variation of temperature on the land are not so excessive as to reverse the direction of the wind, winter and summer, but only to modify its course. Thus, in the central and eastern United States, the prevailing westerlies veer from south and southwest in summer when the continental interior is unduly warm, to west and northwest when it is unduly cold. In Europe, where the greater continental area lies to the east, the monsoon effect is reversed, and there is a change from west or northwest winds in summer to southwest winds in winter. As a compensation for these oblique courses of the surface winds, the upper currents over the eastern United States, as determined by the drifting of lofty clouds, move more from the northwest in summer and from the southwest in winter; while over western Europe their movement is from the southwest in summer and northwest in winter. The contrasted monsoon influences on the two sides of the North Atlantic are of great importance in determining the climatic features of the two regions. They are particularly effective in increasing the annual range of temperature on our eastern coast, and in diminishing it on the western coast of Europe.

156. Continental obstruction of terrestrial winds. Besides introducing complicated variations of temperature, the continental masses obstruct the free passage of the winds. One of the most manifest effects of the inequalities of the land surface is found in the general decrease of the velocity of the winds over the land as compared with that of the winds at sea. The average velocity of the latter approaches twenty miles an hour; while that of the former is not more than half of this value.

The interference with the regular course of the terrestrial winds exercised by the stronger reliefs of the continents is well seen in the western hemisphere, where the Cordilleras of North and South America interrupt the free passage of the winds between the Pacific and the Atlantic. The westerlies of the North Pacific branch southward and join the trades along the coast of California and Mexico; and similarly the trades of the Atlantic give forth a branch that turns northward and reinforces the deficiency in the westerlies in the Mississippi basin east of our Cordilleras. The same thing may be noticed in perhaps better development in the southern hemisphere, where the mean height of the mountain chain is greater than with us. The westerlies of the South Pacific give forth a great branch that turns north along the coast of Chile, and joins the southeast trade off Peru; the southeast trade of the South Atlantic becomes an easterly and northeasterly wind over Brazil, and swings around to join the deficient westerlies in the southern Argentine country. These deviations from the normal paths of the terrestrial winds will be found

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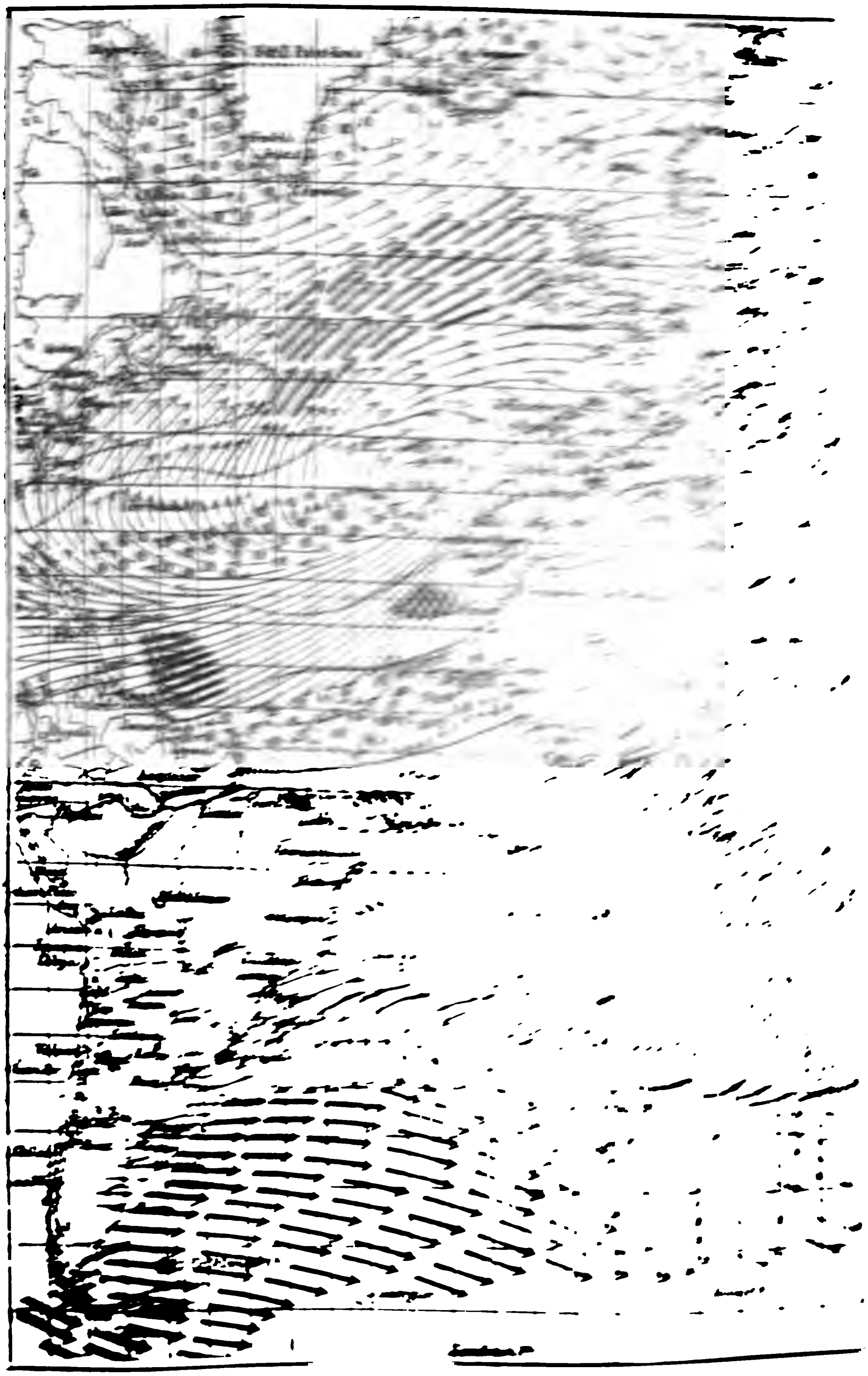
certain members of the terrestrial system being greatly increased by the action of the land in causing a strong migration of the heat equator. In most parts of the world, however, the effects of the variation of temperature on the land are not so excessive as to reverse the direction of the wind, winter and summer, but only to modify its course. Thus, in the central and eastern United States, the prevailing westerlies veer from south and southwest in summer when the continental interior is unduly warm, to west and northwest when it is unduly cold. In Europe, where the greater continental area lies to the east, the monsoon effect is reversed, and there is a change from west or northwest winds in summer to southwest winds in winter. As a compensation for these oblique courses of the surface winds, the upper currents over the eastern United States, as determined by the drifting of lofty clouds, move more from the northwest in summer and from the southwest in winter; while over western Europe their movement is from the southwest in summer and northwest in winter. The contrasted monsoon influences on the two sides of the North Atlantic are of great importance in determining the climatic features of the two regions. They are particularly effective in increasing the annual range of temperature on our eastern coast, and in diminishing it on the western coast of Europe.

156. Continental obstruction of terrestrial winds. Besides introducing complicated variations of temperature, the continental masses obstruct the free passage of the winds. One of the most manifest effects of the inequalities of the land surface is found in the general decrease of the velocity of the winds over the land as compared with that of the winds at sea. The average velocity of the latter approaches twenty miles an hour; while that of the former is not more than half of this value.

The interference with the regular course of the terrestrial winds exercised by the stronger reliefs of the continents is well seen in the western hemisphere, where the Cordilleras of North and South America interrupt the free passage of the winds between the Pacific and the Atlantic. The westerlies of the North Pacific branch southward and join the trades along the coast of California and Mexico; and similarly the trades of the Atlantic give forth a branch that turns northward and reinforces the deficiency in the westerlies in the Mississippi basin east of our Cordilleras. The same thing may be noticed in perhaps better development in the southern hemisphere, where the mean height of the mountain chain is greater than with us. The westerlies of the South Pacific give forth a great branch that turns north along the coast of Chile, and joins the southeast trade off Peru; the southeast trade of the South Atlantic becomes an easterly and northeasterly wind over Brazil, and swings around to join the deficient westerlies in the southern Argentine country. These deviations from the normal paths of the terrestrial winds will be found



are northerly or northeasterly winds issuing from certain parts of the region around the pole; these may be regarded as lingering representatives of a vast system of polar winds that would sweep towards the equator if a strong high pressure at the poles had not been so nearly reversed to low pressure by the whirl of the circumpolar winds. In the southern hemisphere, southeast winds are reported as prevalent in high latitudes: hence an increased pressure may be expected around the pole; but the reduction of pressure around the south pole is more strongly marked and the winds there are stronger than in the northern hemisphere, even though the poleward temperature gradients in that hemisphere are somewhat weaker than in ours. This want of symmetry in the wind systems of the two hemispheres must be referred to the unsymmetrical distribution of land with respect to the equator.

159. Diurnal variation in wind velocity on land. Over the oceans, the velocity of the wind shows no distinct diurnal period; but over the lands and particularly in clear warm weather, the winds are distinctly stronger about noon than in the night. This is fully accounted for by the convectional interchange in the day-time between the surface air and the faster moving currents at a height of one or several thousand feet, as explained in Section 54. At night, the wind near the surface of the earth is greatly retarded by friction, and before morning generally falls to a calm, unless urged by some temporary stormy disturbance: at a height of a thousand feet, the movement of the air continues about as usual. Intermediate layers of air retain velocities dependent on the greater or less frictional resistances that they suffer. But as soon as the instability of morning hours causes convectional interchange between the various layers, their movement is more nearly equalized and hence the velocity of the wind is increased at the earth's surface, reaching a maximum in the early afternoon. As the ground cools towards sunset and the convectional currents cease, the surface winds weaken and the evening becomes calm. It will be seen that the diurnal increase of cumulus clouds (Sect. 196) is caused by the same convectional process. In fair summer weather, the variation in the velocity of the wind and in the amount of cloudiness is very distinct even to ordinary observation. The change in the velocity of the wind is most pronounced in arid regions, such as our drier western plains, where the diurnal variation of temperature in the lower air is strong: at night the calm is often complete; in the day-time the wind may rise to a dusty gale. The dust whirlwinds of desert plains should be associated with this feature of the general winds, as they are only the more visible manifestation of the process on which the diurnal increase in velocity generally depends. The greater velocity of the summer than of the winter monsoon in India, and in general the greater velocity of the wind on a given gradient in summer than in winter, is best explained by this process.

It must be inferred from this peculiarity of the general winds that the weak gradients on which they depend, illustrated in the isobaric charts IV, V, VI, are not sufficient to drive the lower air across the rough surface of the lands, unless aided by the intermixture of the lower and higher layers by convection. When winds occur at night on land, they are to be referred to local gradients, such as are described in the following sections, or such as accompany the passage of storms.

Furthermore, according to this theory, a necessary consequence of the diurnal variation in the velocity of the surface wind over the land is an inverse variation in the velocity of the upper wind. The velocity aloft should decrease by day, for the gravitative acceleration on the general gradients has then to expend a part of its force in overcoming friction with the ground; and the velocity aloft should increase at night, when the lower air lies still and the acceleration has to overcome only the friction of air on air. The records of mountain observatories show very clearly that the variation thus called for by theory really exists. Even at the moderate height of the Eiffel tower (990 feet) in Paris, the nocturnal increase in the velocity of the wind is distinct. The explanation as applied to the lower winds is due to Espy, who announced it in 1840; the variation of the upper winds was detected by Hellman of Berlin, in his study of our Mt. Washington records in 1875; it was explained by Köppen in the same year.

The vertical interchange of upper and lower air in the day-time has been applied to explain the hot noon-time westerly winds of the plains of northern India. These winds have a less amount of moisture than is observed in the region whence they seem to come. It is ingeniously suggested that this peculiarity results from a convectional descent of dry upper air from an elevation of about 10,000 feet, to which the surface heat would cause the lower air to ascend. At that level, the calculated gradients would cause westerly winds; and the dry upper air, descending, would reach the ground with an abnormal direction, a high temperature, and an unusually small amount of moisture. A similar explanation may be found to apply in northern Texas and Kansas where parching hot westerly winds are known on summer days (Sect. 245).

There is a slight *diurnal variation in the mean direction of wind,* of small practical importance, but of much interest from the evidence that it gives of the correctness of the principles in accordance with which the theory of the winds has been framed. The average of many hourly observations shows that the wind tends to veer a little to the right in this hemisphere as the day passes, and to turn back again as night comes on. This is in consequence of the convectional descent of the upper air, as just explained; for the upper currents, with little friction, turn strongly from the gradient, and this effect is propagated downward in the day-time towards the earth's surface; but at night

when surface friction with the ground is more largely in control, the deflection from the gradient is less; hence the surface wind turns a little to the right of its mean direction by day, and to the left by night.

160. Land and sea breezes. The seasonal contrasts of temperature on land and water have a parallel in the diurnal contrasts. In summer particularly, the land over its whole area is warmer than the sea by day, and cooler by night; and if our days were long enough, diurnal winds would sweep into the continents to their very centers every day, and back again every night. But the rotation of the earth is so rapid in comparison to the circulation of the winds excited by the diurnal contrasts of temperature that there is not time for their motion to be propagated far inland or seaward from the coast-line. The litoral breezes seldom extend more than twenty or thirty miles inland, and their seaward extension is probably less. Delicate observations have detected a slight diurnal variation of pressure at stations on the coast and over the neighboring interior, similar to that between the sea and land in winter and summer, but of much less amount.

The sea-breeze begins in the morning hours, from nine to eleven o'clock, as the land warms; it brings in the pure cool air from the sea. In the late afternoon it dies away, and in the evening the land breeze springs up and blows gently out to sea till morning. This process is repeated with great regularity in the tropics, where there are few storms and the diurnal changes are the chief ones. In our latitudes, the land and sea breezes are often masked or overcome by the winds of cyclonic storms; they appear only in the spring or summer, when the air over the land may become for some hours decidedly warmer than over the sea. The sea breeze reduces the temperature on its arrival and prevents a high noon maximum; a thermograph often exhibits an early morning and a late afternoon maximum, with a moderate depression in the curve between the two during the blowing of the breeze, as appears in Fig. 11. It thus diminishes the range and lowers the mean temperature on the coast. On tropical coasts, the sea breeze is the healthful wind; the land breeze, often blowing from miasmatic swamps, is fever-laden, bearing the odors of the soil (Sect. 310).

The changes of pressure of diurnal period on the Iberian peninsula may be here referred to, in continuation of the illustrations of annual variations of temperature and pressure already afforded by that region. The mean pressure for 9 A.M. for July, Fig. 42, exhibits a weaker system of centripetal gradients than appears for the mean of July, as given in Fig. 37; but the pressure for 3 P.M. for July, Fig. 43, shows gradients of greater value. In this case, we may therefore expect that the occurrence of the litoral sea breeze of day-time is associated with an increase of the general inflow of the summer winds; while in the early morning, before the temperature has risen,

the inflow must be weaker. A figure for nocturnal hours has not been prepared, for lack of observations; it would probably show a further weakening of the inflow, with an off-shore gradient around the coast.

The land and sea breezes follow the local gradients closely when they begin; but as they are drawn in from a greater distance, they manifest a distinct tendency to deflection. If the sea breeze is easterly at first, it veers

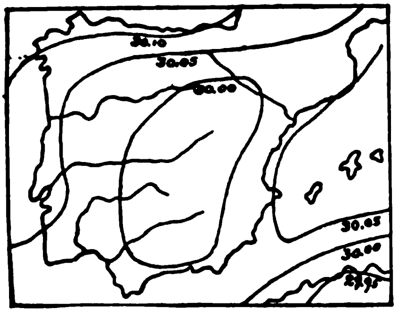


FIG. 42 (July, 9 A.M.).

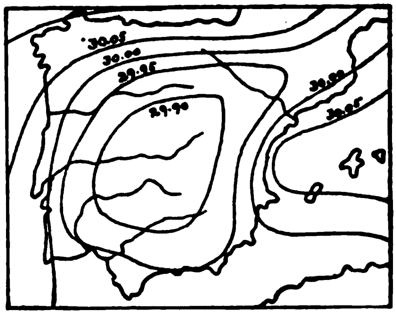


FIG. 43 (July, 3 P.M.).

to southerly before fading away; then the land breeze, coming first from the west, veers toward the north late at night. Thus a systematic diurnal rotation of the wind is produced; such veering breezes are locally known as "roundabouts" on the coast of Massachusetts. The same change is well known elsewhere. It is often said: we know the earth turns around, because the sun and stars rise and set. It might also be said: we know the earth turns around, because the trades and the westerlies blow obliquely, or because the land and sea breezes veer in regular order.

The average hourly direction of the land and lake breeze at Chicago for July, 1882, Fig. 44, gives excellent illustration of a regular veering from the lake breeze of afternoon into the land breeze at night, with a sudden reversal

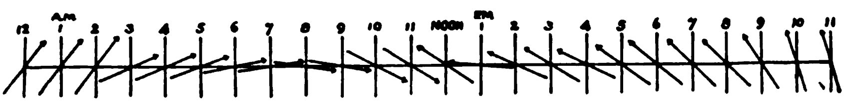


FIG. 44.

of the latter into the former at noon. The sea breeze is felt earlier on low ground than on high; the depth of the breeze on our sea-coast may be roughly gauged from observations made in a captive balloon at Coney Island, near New York; the average elevation at which the cool inflow from the sea was exchanged for the overlying warm outflow from the land was from five to six hundred feet.

161. The sea breeze begins off-shore. There is one peculiarity of the sea breeze that deserves mention, as it does not appear in the continental winds whose period is so much longer than twenty-four hours. The sea breeze, as a rule, makes its first appearance off-shore and gradually beats its way to the land. It would appear at first sight that its growth should be in just the other direction; for when the pressure decreases over the land by the overflow of warmed air, there should be an immediate response from the adjacent lower air, which, it would seem, should at once flow in to the region of diminished pressure, and then gradually extend its area backwards to a greater and greater distance from the land. But such is not the case; and the following explanation has been offered by Seemann to account for the observed facts. The winds that have been thus far explained are examples of steady motion, in which a condition of equilibrium has been attained between the accelerating force of gravity, the deflective force of the earth's rotation and the resistance of friction (Sect. 94). Let us suppose for a moment that in the case of the land and sea breeze, the change of temperature in the air over the land is immediate, warming at once at sunrise and cooling as quickly at sunset. There would manifestly be a time following these abrupt changes in which the circulation of the air was not adjusted to the gradients offered to it. The rapid expansion of the air on the ground could not wait for the overflow of the air from above it, but would at once demand and secure more space for itself by expanding laterally to seaward, where the air is cool and its expansive force relatively weak. But as the upper air flows away and relieves the lower air from its constraint, the conditions of ordinary convectional circulation gradually appear. These conditions will be established first a short distance from the shore where the overflow from the land accumulates, and will gradually make their way to the land.

The actual case has not the immediate changes of temperature here supposed, but it may be presumed that the rate of warming of the air on the land is for a time in the morning so rapid that the expansion of the air keeps in advance of the overflow by which it is accommodated; and as long as the rise of temperature thus causes an increase in the gradients, steady motion cannot be reached. A reaction, such as the pressure of the land air toward the sea, must exist as long as the gradients and the rate of motion on them are increasing; the delay in the establishment of the sea breeze and its first appearance in the offing may be ascribed to this reaction. A reported increased strength and high temperature of the tropical land-breeze in the morning, just before the sea breeze sets in, is confirmatory of this theory.

162. Combination of general and litoral winds. When the changes of temperature that would cause a normal sea breeze in a stationary atmosphere take place in a wind of other origin, they modify its direction or intensity.

The westerly wind that prevails in Ohio in summer is deflected in the neighborhood of Lake Erie into a northwesterly wind by day and a southwesterly wind by night, because of the contrasts of temperature on either side of the shore-line, which lies about parallel to the course of the general wind. On the northern coast of Long Island Sound, in fine summer weather, the wind generally comes from the west, but shifts from northerly at sunrise to southerly at noon. On the coasts of California and Chile, the prevailing west wind is intensified in summer time into a moderate gale by day and retarded to a calm or even reversed to a light land breeze by night. The east coasts in the trade-wind belt show a similar variation in the strength of their winds.

163. Mountain and valley breezes. The seasonal and diurnal winds that have been considered thus far would appear on a level earth. There remains a class of breezes of diurnal period whose opportunity depends on the unevenness of the land surface. These are felt in valleys at night, blowing down stream and increasing to a brisk gale where a large valley emerges on an open plain; and on the higher mountain sides by day, blowing up the slope, but with less velocity than is attained by the concentrated nocturnal down-flow.

The explanation of the nocturnal mountain breeze is not far to seek. It results simply enough from the faster cooling of the surface stratum of air that has already often been referred to; when the cool and therefore heavy stratum lies on a slope, it causes descending currents. If the form of the surface concentrates this aërial drainage from a large upland region into a narrow valley outlet, a mountain breeze of some violence will appear in its

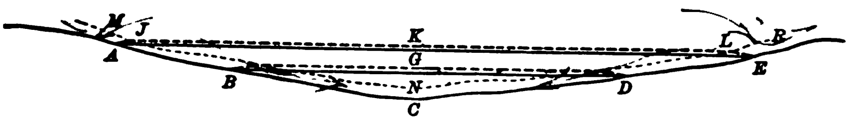


FIG. 45.

lower course. The up-stream valley breeze of day-time may be explained as follows: Let *ABCDE*, Fig. 45, be the cross-section of a valley, in which the morning air lies quiet, with level isobaric surfaces, such as *AGE*. As the day advances and the lower air, *MNR*, warms, it expands and lifts the overlying air bodily. The greater part of the isobar will be evenly lifted by the amount of the expansion in the air near the surface beneath it, and will be found at *JKL*; but towards the sides of the valley, where the isobar runs near the ground, the lifting that it suffers is less, and at the points *A* and *E* its position remains unchanged. At either side of the valley faint gradients, *JA* and *LE*, are thus formed, on which the air is urged towards the slopes, and there, by reaction with the ground, an up-hill current is formed. As in the previous

case, it appears with greater strength where the form of the surface concentrates its flow, and hence is to be looked for in lateral valleys rather than on the spurs between them. As the temperature of the ascending air is lowered by expansion, the valley breezes that rise past a mountain peak may retard its diurnal rise of temperature, as seems to appear in Fig. 12, b.

Fragmentary accounts of such winds are found in the narratives of our western exploring expeditions, but they have not yet received the careful description that they deserve. They are doubtless to be met with in all parts of the western mountainous country; the deep valleys leading out from the Sierra Nevada into the plain of California should show the nocturnal mountain breeze to perfection. In the Andes and the Himalaya they are well known; in the latter they are described as blowing up the valleys by day from nine o'clock in the morning to an early hour in the evening; at night they blow down again, and where the larger streams open out to the plains, they attain some violence: but on the high plateaus the nights are calm.

In elevated valleys between snow-covered mountains the air lying on the snowy slopes remains cold through the day, while the air on the valley ground is warmed; in such cases a cold, stormy descending current is felt on the slopes, and an ascending current may be expected over the middle of the valleys. The lateral descending current has been observed on the high snow-fields of the Andes. On the other hand, when a valley is occupied by a glacier, the air near the ice is held at a low temperature, and may, even in day-time, form a mountain breeze, which is ordinarily limited to the night. Descending glacial breezes of this kind are reported from the Muir glacier of Alaska.

The presence of mountains near a coast-line where the land and sea breezes are felt, serves to intensify them by the additional causes of motion then introduced. Even the great monsoon system of Asia is aided in this way, Asia being a land of great relief.

164. Mountain breezes and inversions of temperature. The nocturnal inversions of temperature, described in Section 43, characterize clear weather on plains, where the wind falls to a calm in the evening. There can be little question that balloon ascents over plains during clear, quiet nights would in nearly all cases discover an increase of temperature with ascent for several hundred feet. This inversion of the vertical temperature gradient has been ascribed to the cooling of the lower air by conduction and radiation to the cooled ground; it is independent of any motion in the air.

In hilly regions the nocturnal inversions of temperature are not only rendered more apparent from the ease with which they may be observed; they are intensified by a gentle aërial drainage down the slopes, analogous to the mountain breezes above described, but of much gentler motion. As night comes on, the chilled air on the hill-tops and slopes runs down into the

adjacent valley, and there accumulates in much greater thickness than it would gain on level ground; while the hill-tops continually receive a supply of uncooled air, settling on them from above, and retain a relatively high temperature. The marked contrast between the mild air of midnight on the hills and the chilly air in the valleys may often be observed even where the difference of level does not reach a hundred feet.

In the case of a strong mountain breeze issuing from a large valley out upon a broad plain, a somewhat different result may be expected. As long as the descent of the breeze is slow, the loss of heat by conduction and radiation to the ground may entirely overcome its tendency to an adiabatic increase of temperature by compression in descent. If not counteracted, this would cause a rise of one degree for every 188 feet of descent. When the descent is rapid, as in a breeze draining a large surface of mountain slopes, all converging to discharge by a single valley outlet, then the gain of heat by compression may appreciably decrease the cooling of the breeze; and as such a breeze issues from the valley-mouth to the plain, it may be of higher temperature than that to which the quiet air at the same level on the plain some distance away has fallen. While this is in a measure only a theoretical deduction, it receives some confirmation from an account of a western military station at the foot of a mountain range where a valley opened to an elevated plain. The height of the plain was too great for the successful growth of wheat; but near the fort, by the mouth of the valley, wheat was safely harvested. Although the observers there did not mention the mountain breeze as the local safeguard against nocturnal frosts, it does not appear unlikely that such may have been the case.

165. Winds not yet classified. If the charts of the average winds for January and July are now examined, it will be found that nearly all of the many directions that the winds follow in one region or another may be explained by reference to some of the foregoing paragraphs. But if our daily weather-maps are examined, we may often see south winds in the Mississippi valley in winter, and north winds there in summer, which find no explanation from the causes thus far considered. Their origin will be stated in the chapter on cyclonic storms: but as these storms are always accompanied by clouds and rain, some account of the moisture of the atmosphere must be next introduced.

CHAPTER VIII.

THE MOISTURE OF THE ATMOSPHERE.

166. Evaporation. The presence of a water surface over three-fourths of the earth insures a continual supply of water vapor for the atmosphere; while the interruption of the ocean surface by continents and the great variations of temperature in time and place require that the quantity of vapor in the atmosphere shall continually vary. To appreciate its changes we must examine the process of evaporation; the distribution of the vapor through the atmosphere; and the processes by which it may be condensed again into the liquid or solid state.

The process of evaporation or the change from the solid or liquid to the gaseous state requires the expenditure of a large amount of energy. Liquefaction or the change from the solid to the liquid state also requires a considerable supply of energy, but with this process we are not so much concerned. The melting of ice and snow must be duly considered, but the evaporation of water is of greater importance in meteorology.

It is supposed that the energy needed in evaporation of water is expended in overcoming the attraction that exists between the molecules while the water is in the liquid state. The supply of energy to do this hidden work often comes from the sensible heat of some adjacent substance. When water evaporates from the sea or from a lake or river or from the wet surface of the land, the energy needed to change its state may be derived in part from the heat of the adjacent water or land; but in the usual case of evaporation proceeding under sunshine, it is supposed that the energy of insolation may pass directly to the work of overcoming the intermolecular attractions of the water and thus changing it to the gaseous state, without taking the intermediate form of heat. This is illustrated in the ordinary experience of a drying day after a rainstorm. The surface of the land, everywhere wet from the rain that fell from the clouds of the day before, is then shone upon by the sun's direct and indirect rays from the clear sky. Instead, however, of there being a rapid rise of temperature, there is a rapid drying of the ground; the energy of insolation received upon the surface of the ground is expended in changing the state of the water more than in increasing the molecular activity of the ground or of the water. In the same way, the strong insolation absorbed at the surface of the torrid oceans is devoted more to causing evaporation than to raising the temperature of the water; hence in good part for this reason the oceans around the equator are relatively cool.

Water vapor is lighter than the other gases of the atmosphere. The weight of a cubic foot of vapor at a given temperature and under a given pressure is only 0.630 of the weight of the same volume of air under the same conditions. A cubic foot of moist air is therefore lighter than a cubic foot of dry air at the same temperature and pressure; but when water evaporates into a given volume of air, the weight of the mixture and its expansive force are both increased by the presence of the vapor. As vapor is formed from a wet surface, it spontaneously but rather slowly mixes with the air; this process being called diffusion. The distribution of vapor through the atmosphere is, however, chiefly controlled by the movement of the air in its local and general circulations.

167. Latent heat. The energy required to change a substance from the solid to the liquid state, or from the liquid to the gaseous state, is called *latent heat.* This term is essentially a misnomer. It was introduced into the science of physics at a time when very different ideas concerning the nature of heat prevailed from those that are now current. It was then thought that heat was an imponderable form of matter, and that this imponderable matter had to enter any liquid, as water for example, in the process of evaporation; but as evaporation is not attended necessarily by a rise of temperature, the heat thus entering the water was said to become *latent.* As now understood, heat is not a form of matter, but a condition of matter, and latent heat is not heat, but is simply the energy needed to overcome the intermolecular attractions of the evaporating substance; yet the name, *latent heat,* is still universally employed. The student must be careful to avoid being misled by its apparent meaning. It is simply a special name for the energy, from whatever source, expended in a certain task; its addition does not cause any change of temperature whatever.

The amount of energy required to melt a pound of ice at 32° would raise a pound of water from 32° to 172°. If a pound of fresh snow at a temperature of 32° were placed in a pound of water at 172°, the resulting two pounds of water would have a temperature of 32°. Hence the latent heat of liquefaction of a pound of ice is equal to 140 units of heat,—a unit of heat being the amount needed to raise the temperature of a pound of water one degree Fahr. The importance of this in explaining the low polar temperatures in summer under a strong supply of insolation has already been referred to in Section 91.

The amount of heat required to evaporate a pound of water is much greater than that required to melt a pound of ice. About a thousand units of heat are needed to transform a pound of water into a pound of vapor; the precise amount varying with the temperature at which evaporation takes place. At 32° it is 1092 units; at 212° it is 966. The latent heat of water vapor is therefore very high.

With this in mind, we may recall the small diurnal change of temperature in the surface waters of the ocean. Even under the strong shining of an equatorial sun, the surface of the sea in the torrid zone warms but two or three degrees from night to day, while land surfaces in the same latitude may warm by twenty or thirty degrees. The share of insolation that is absorbed at the ocean surface goes for the most part, not to exciting the molecules of the water to faster motion, that is, to heating the water, but to the other task of changing the state of the water from liquid to gas, that is, to supplying the latent heat needed for evaporation from the water surface.

168. Capacity. The quantity of vapor that can exist in the air depends upon the pressure that is exerted upon it and upon its temperature; but it is important to notice that the controlling pressure is only that which is exerted by the vapor itself, and not by other gases with which it may be mixed. Thus at a certain temperature, as 80°, such as occurs frequently in the lower air over the torrid oceans, the vapor may increase in quantity until its expansive force equals about one inch of barometric pressure. If the total pressure is then thirty inches, the expansive force of the lower air with which the vapor is mixed will be twenty-nine inches. Hence, although the temperature of the air determines the temperature of the vapor and thus controls its amount, the pressure of the air has no effect in determining the quantity of vapor that may be formed, although it is important in diminishing the rate of evaporation, because it retards the rate of diffusion through the air. It is as if the molecules of the other gases of the atmosphere acted only as so many obstacles which the molecules escaping from the water surface had to pass by; for the total quantity of vapor that could be formed at a temperature of 80° would be just as great if the air were absent as in its presence. Inasmuch, however, as the air, into which evaporation proceeds, in nearly all cases determines the temperature of the vapor, it is natural and usual to speak of the *capacity of the air for vapor.* This is really a misleading expression, like many others inherited by science to-day from earlier years, but its convenience warrants its adoption.

169. Saturation. When the full capacity of a given volume of air for vapor has been reached, the air is said to be *saturated* with vapor. A more careful statement of this condition would be given in the phrase, the vapor is saturated; for it is believed that, if any additional evaporation should take place under such conditions, some of the preëxistent vapor must return to the liquid state. This phrase, however, is seldom employed; it is sufficient to say that the air is saturated. This state is therefore spoken of as saturation.

The capacity of air for vapor increases rapidly with rise of temperature. At a temperature of zero, Fahrenheit, the expansive force of the greatest

quantity of vapor that can then exist can be no more than 0.04 inch; at freezing it is 0.19 inch; at 90°, 1.41. The following tables exhibit this relation more fully, and add also the maximum weight of vapor in grains per cubic foot, and in grams per cubic meter, at various temperatures; and the weight of a cubic foot of saturated air under a pressure of 30 inches.

PRESSURE AND WEIGHT OF VAPOR AND SATURATED AIR.

TEMPERATURE. °F.	VAPOR PRESSURE. *Inches.*	VAPOR WEIGHT, CU. FT. *Grains.*	SAT. AIR WEIGHT, CU. FT. *Grains.*
− 30°	0.010	0.12	650
− 20	.017	0.21	634
− 10	.028	0.35	620
0	.045	0.54	606
+ 10	.071	0.84	593
20	.110	1.30	580
30	.166	1.97	568
40	.246	2.86	556
50	.360	4.09	544
60	.517	5.76	533
70	0.732	7.99	521
80	1.022	10.95	509
90	1.408	14.81	497
+ 100	1.916	19.79	487

TEMPERATURE. °C.	VAPOR PRESSURE. *mm*	VAPOR WEIGHT, CU. MET. *Grams.*	SAT. AIR WEIGHT, CU. MET. *Kilogr.*
− 30°	0.38	0.44	1.45
− 20	0.94	1.04	1.40
− 10	2.15	2.28	1.35
0	4.57	4.87	1.30
+ 10	9.14	9.36	1.25
20	17.36	17.15	1.20
30	31.51	30.08	1.15
+ 40	54.87	50.67	1.11

A room measuring twenty feet square by ten feet high would contain 4,000 cubic feet of air. If it were saturated with vapor at a temperature of 60°, the weight of the moist air would be 304 pounds avoirdupois; and the vapor would weigh 3.3 pounds. If all the vapor were condensed, it would produce 91.4 cubic inches of water, or somewhat more than three pints.

CHAPTER IX.

DEW, FROST AND CLOUDS.

180. Condensation. The natural processes of condensation of the water vapor in the atmosphere all depend on a decrease of temperature. It is possible in the laboratory to produce condensation by the compression of moist air, the temperature in the meantime being maintained at a constant value; but this process has practically no application in meteorology; for when air containing vapor is compressed, as in a descending current, the diminution of capacity resulting from decrease of volume is more than made up by the increase of capacity resulting from increase of temperature; and descending currents for this reason soon become dry.

Condensation as a result of cooling the air has already been briefly referred to in Section 172. Any mass of air containing vapor will, if cooled sufficiently, be reduced to the dew-point, and if the cooling then proceeds further, progressive condensation will accompany it. If condensation occurs at temperatures above 32°, the product will be water; if below 32°, it will be ice in the form of frost, snow or hail.

181. Condensation from quiet air on cold surfaces. There are various processes by which the temperature of the air is cooled to the point of condensation. One of the simplest of these is illustrated in the cooling that accompanies the diurnal changes of temperature in the atmosphere. These have been already explained to be small in the upper air and greatest near the ground, and the latter occurrence will therefore be first considered. As the temperature rises in the morning, any moist surface rapidly yields its vapor to the warming air. As a rule, however, the increase of temperature goes on so rapidly that the supply of vapor does not cause saturation. During the morning hours, even though the absolute humidity increases slightly, the relative humidity quickly falls. During the first hours of the afternoon, if evaporation continue from the warmed ground into the warm air, the absolute humidity still increases slowly, but at this time the temperature is falling and the capacity of the atmosphere for vapor is thereby decreasing; hence the relative humidity rapidly increases. About sunset in well-watered regions, the air close to the ground is nearly saturated, as we may know from the growing dampness of the grass; and from this time on the further cooling of the ground during the night and the consequent cooling of the air next to it by radiation and conduction, causes the continuous deposition of vapor in the form of dew or frost; the former if condensation occurs at temperatures above

of clothing; if the air is windy, more protection is needed than when it is calm; if it is damp as well as cold and windy, it abstracts all the more heat from us, probably by means of the better conductivity given both to the air and to the clothing by the moisture; hence the difference between the bracing though severe cold of our dry northwest winter winds, and the penetrating, searching chill of our damp winter northeasters. The difference between the so-called "dry cold" of the interior and the "damp cold" of the New England coast is thus explained. On the other hand, when the air is warm, our bodily temperature would rise too high if it were not for the cooling of the skin by continual evaporation from its surface. In very hot and very dry air, the evaporation is so much hastened that the skin is parched and burned; in hot and very damp air, evaporation is checked and the air feels sultry and oppressive. Moderately dry hot air is less uncomfortable than at either of the extremes of dryness or dampness. The oppressiveness of our "dog-day" weather in July and August depends as much on its humidity as on its heat.

The action of water vapor on insolation and terrestrial radiation has been much discussed. It may be regarded as diathermanous to insolation, but relatively opaque to terrestrial radiation, and it is therefore thought to exert a controlling influence in determining the temperature of the atmosphere. It is for this reason, as well as for the reasons stated on pages 29 and 33, that the range of temperature is large in arid regions, and small over the oceans. Without water vapor, the temperature of the earth would probably be much lower than that now prevailing. This appears to be confirmed by observations on the diurnal range of temperature under varying conditions of humidity. If the temperature of the air is well above saturation, the range is relatively strong; if near saturation, the range is diminished even though no visible clouding of the sky occurs; if a thin, hazy cloud is formed, the range is greatly reduced.

171. Absolute and relative humidity. In order to measure the relative dryness or dampness of the air, it is customary to determine the ratio of the amount of vapor actually present to that which might be present at the existing temperature. The amount of vapor actually present is called the *absolute humidity*. This may be expressed either in the expansive force that the vapor exerts or in its weight in grains per cubic foot of air. The absolute humidity divided by the amount of vapor that might exist if the air were saturated gives a ratio that is called the *relative humidity*. Close over the ocean surface, the relative humidity is generally over 90 per cent, and may reach, at night, 100 per cent; that is, the condition of saturation. In our dry winter weather, the relative humidity may fall below 50 per cent, and sometimes as low as 40 or even 30 per cent. In deserts, 20 per cent is not uncommon; and at noon-time, when the rapid rise of temperature gives the air a greatly increased capacity for vapor which cannot be satisfied, the relative

the surface of the ground is dried; water rises from the subsoil by capillarity to supply more vapor to the thirsty air. The sap passing from the leaves of plants is also easily disposed of then, because it is exuded in the form of extremely minute drops, and because it is generally well exposed for evaporation to the surrounding air. But at night, drops of exuded water may collect on the leaves of grass and low plants, where it is unable to evaporate in the cold nocturnal air; and water rising to the surface of the bare ground may remain there, instead of passing off as vapor. If the ground is covered with grass, the blades of grass become colder than the ground beneath them; the vapor rising from the ground will then be in good part condensed on the cold grass.

There is a large variation in the proportion of dew supplied from its several sources at different times and places. In damp countries, the share coming from the ground is large; in dry regions, where the dew formed at night may be a large part of the water on which the growth of plants depends, the greater part of it presumably comes by condensation from the air. Few careful observations have yet been made on this subject.

184. Frost is formed in the same manner as dew, but at temperatures below the freezing point. The frost that forms just below the surface of the ground, raising the surface soil by the growth of its crystals, is probably derived for the most part from the subsoil: that which is deposited on loose leaves and sticks lying on the ground may also be supplied in good part from the ground; the share that then comes from the air is not yet determined.

Little attention has been given to the measurement of the diurnal or annual amount of dew and frost, on account of the difficulty of the problem. It has been estimated that the dew of a single night equals 0.02 inch of rainfall (Sect. 283); and that the total annual amount of dew precipitated in Great Britain would measure an inch and a half in depth. Further study should be given to this subject.

185. Conditions for the formation of dew or frost. In dry regions, even when the range of temperature is great and the nights are much cooler than the days, it may be that the dew-point is not reached at night and no dew is formed. This is commonly the case in deserts; but in regions of moderate aridity, where the rainfall is of small quantity and the days are warm and dry, the nights may often be cool enough to form dew or frost, as on our western plains, and on the pampas of the Argentine Republic. In well-watered regions, such as the eastern United States, the formation of a moderate amount of dew or frost in favorable situations is common on all clear and quiet nights, and it is accompanied by inversions of temperature of greater or less strength. The damp air of the torrid zone yields abundant dew on many equatorial lands

ground is increased by plowing or otherwise breaking up its surface. Plants exude water from their leaves in the growing season; deep-rooted trees and grasses bring a large amount of ground water up to the air in this way. Evaporation is active from high mountain surfaces in clear weather in spite of their low temperature, because of the open exposure of their surface to the dry and active currents of the upper atmosphere.

The amount of evaporation that may take place from a free water surface continually exposed to the air has been determined for various parts of the world. The measures are not closely comparable, because they come in some cases from the loss by evaporation from large reservoirs, in which the temperature of the water may differ from that of the air and thus exercise a large control on its evaporation; and in other cases they are determined by the loss from comparatively a small volume of water in a shallow vessel, whose temperature may follow that of the air within a few degrees. The former method is more useful in connection with engineering works, such as reservoirs for supplying cities, or for storing water to be used in irrigation. At inland stations of dry regions the amount thus determined does not correspond to any natural quantity, but in the neighborhood of the sea or of large lakes it serves to measure roughly the amount of water that passes from their surface into the atmosphere. At stations near the continental coasts the amount of evaporation is generally a little less than the annual rainfall; but there must be many local exceptions to this rule. The records for a number of stations are here given; those for our western interior basins are general averages.

ANNUAL EVAPORATION FROM FREE WATER SURFACES IN INCHES.

PLACE.	LATITUDE.	EVAPORATION.
Madras	13 N.	91.2
St. Helena	17 S.	83.8
Dijon, France	47 N.	26.2
London	51 N.	20.6
Boston	42 N.	39.1
Lake Michigan	44 N.	22.–
Great Salt Lake	41 N.	80.–
Interior Basin, U. S.	36 N.	150.–
Interior Basin, U. S.	44 N.	60.–
Ft. Conger	82 N.	8.9

174. Hygrometry. Hygrometers are instruments for measuring the humidity of the air. The simplest instruments of this kind employ some hygroscopic substance, — that is, one which easily absorbs moisture from damp air, — such as a hair from which the natural oil has been extracted by placing it in an alkaline solution. The hair is fastened at one end, passed over an easily rotating cylinder, and held tense by a weight or spring at the other

condensation begins at temperatures above 40°, it is seldom that the minimum temperature of the night will fall to freezing; but no definite rule can be given in this case. Local experience is needed to determine the "safety limit" of the dew-point and its relation to the season and the weather. Special study might be profitably turned in this direction.

187. Protection from frost. When the occurrence of frost appears likely, it is often possible to protect plants from injury by building a smoky fire on the windward side of a field, so that a dense stratum of smoke may drift slowly over the surface. Radiation is then transferred in great part from the ground and the plants to the smoky stratum, and the injurious fall of temperature at the level of the ground is effectively retarded. This method is often practised to protect tobacco in meadows and cranberries in low boggy ground, where the air is more quiet than on hills or slopes. On higher ground such protection is less often needed; for the more active movement of the air over hills, coupled with the conditions of nocturnal inversions of temperature and the nocturnal drainage of the air on sloping ground, generally serves to maintain the minimum temperature in quiet weather on hillslopes several degrees above that experienced in the neighboring valleys and lowlands. Peach orchards in the more northern states should be for this reason generally planted on rising ground: many examples could be given in which the blossoming in an orchard on higher ground escaped freezing, while another near by on lower ground had all its trees blighted. The limitation of low temperatures to the lower layers of the air is sometimes so marked that the upper branches of small trees or shrubs escape a frost that injures the lower branches.

188. Valley and lowland fogs. Valley bottoms are, in spite of the diminished nocturnal cooling of their soil, as a rule much damper at night than the adjacent hilltops; evaporation may continue from the little-cooled surface of their streams and ponds; they may not only receive a plentiful deposit of dew on the ground, but their lowest part may be covered with a layer of mist or fog. Common experience gives many examples of this; when driving late at night over a hilly road, the change of both temperature and dampness from hill to hollow is perfectly apparent. In the torrid zone, where the diurnal changes of temperature are peculiarly regular, the occurrence of nocturnal valley fogs is a characteristic feature of the climate. In the winter of temperate latitudes, when the long nights are interrupted only by short days of weak sunshine, the valleys of mountainous regions may be filled to a considerable depth with a cold damp fog; in spells of quiet and cold winter weather, broad lowlands are sometimes covered by fog sheets of this kind, extending over thousands of square miles, attaining a thickness of a

bulb thermometer, and hence it will read the same number of degrees as its neighbor; but in dry air, evaporation will take place so rapidly from the moist surface of the bulb as to maintain its temperature several degrees lower than that indicated for the air by the other thermometer. The difference between the dry and the wet bulb may then be used with tables constructed from laboratory experiments to determine several quantities desired; the dew-point, the relative and the absolute humidity. The chief difficulty in using the psychrometer consists in determining the proper reading of the wet bulb thermometer, on account of the uneven movement of the surrounding air. Under given conditions of humidity, if the air is stagnant, evaporation from the wet bulb may cause an approach to saturation in the air close about it. Further evaporation is thus retarded, and the depression of the wet bulb reading is reduced. On the other hand, when the atmosphere is of the same humidity as before, but is in active motion, fresh bodies of air are continually carried past the wet bulb; evaporation is much more active, and the depression of the wet bulb reading is decidedly increased. It is for this reason that some standard rate of air-movement past the psychrometer should be maintained. This is sometimes accomplished by creating an artificial draft by means of a fan driven at a constant speed by clockwork. A simpler and much cheaper method consists in attaching the psychrometer to a string a few feet long (Fig. 47), and whirling it, at a velocity of twelve or fifteen feet a second, around the hand. This must be continued in the open air until a constant difference is maintained between the readings of the two thermometers. Results obtained in this way are regarded as trustworthy, although even at best the measures of humidity are of relatively local value. At temperatures near the freezing point, the indications of the hair hygrometer are thought to be more accurate than those of the psychrometer.

FIG. 46.

FIG. 47.

The following table, taken from a much more extended one by Hazen, is adapted to the readings of a *whirled* psychrometer. It gives the dew-point and the relative humidity for different air temperatures and for various

differences between the whirled dry and wet thermometers; it may be used as giving a general indication of the values of these quantities at altitudes less than 3,000 feet. More extended tables should be employed in reducing regular observations.

Difference of Readings of Dry and Wet Bulbs.		Temperature of Air — Fahrenheit.											
		−10°	0°	10°	20°	30°	40°	50°	60°	70°	80°	90°	100°
1	D.P. . . .	−22	− 7	5	16	27	38	48	58	69	79	89	99
	R.H. . . .	55	71	80	86	90	92	93	94	95	96	96	97
2	D.P. . . .	−76	−18	− 1	12	24	35	46	57	67	77	87	98
	R.H. . . .	10	42	60	72	79	84	87	89	90	92	92	93
3	D.P. . . .	–	−39	−9	7	21	33	44	55	66	76	86	96
	R.H. . . .	–	13	41	58	68	76	80	84	86	87	88	90
4	D.P. . . .	–	–	−22	1	17	30	42	53	64	74	85	95
	R.H. . . .	–	–	21	44	58	68	74	78	81	83	85	86
6	D.P. . . .	–	–	–	−18	7	24	37	49	61	72	82	93
	R.H. . . .	–	–	–	16	38	52	61	68	72	75	78	80
8	D.P. . . .	–	–	–	–	−8	16	31	45	57	68	79	90
	R.H. . . .	–	–	–	–	18	37	49	58	64	68	71	74
10	D.P. . . .	–	–	–	–	–	4	25	40	53	65	77	87
	R.H. . . .	–	–	–	–	–	22	37	48	55	61	65	68
12	D.P. . . .	–	–	–	–	–	−16	17	35	49	62	74	85
	R.H. . . .	–	–	–	–	–	8	26	39	48	54	59	62
14	D.P. . . .	–	–	–	–	–	–	5	28	45	58	70	82
	R.H. . . .	–	–	–	–	–	–	16	30	40	47	53	57
16	D.P. . . .	–	–	–	–	–	–	−20	20	39	54	67	79
	R.H. . . .	–	–	–	–	–	–	5	21	33	41	47	51
18	D.P. . . .	–	–	–	–	–	–	–	8	33	50	63	76
	R.H. . . .	–	–	–	–	–	–	–	13	26	35	41	47
20	D.P. . . .	–	–	–	–	–	–	–	−13	25	45	60	73
	R.H. . . .	–	–	–	–	–	–	–	5	19	29	36	42
22	D.P. . . .	–	–	–	–	–	–	–	–	15	39	56	69
	R.H. . . .	–	–	–	–	–	–	–	–	12	23	32	37
24	D.P. . . .	–	–	–	–	–	–	–	–	0	32	51	66
	R.H. . . .	–	–	–	–	–	–	–	–	6	18	26	33

176. Distribution of vapor in the atmosphere. Vapor is distributed through the atmosphere by two processes. One of these is the spontaneous diffusion of vapor, even when the air is calm. It is probably in great part by this process, aided by the increased expansive force of the newly added vapor, that the water evaporated from the torrid oceans in the calm morning air of the doldrums is prompted to ascend to higher levels and form clouds in the afternoon. Diffusion, however, is a slow process compared to the distribution of vapor by the movement and intermingling of the air; and as the general circulation of the atmosphere is continually maintained in a thorough manner,

the vapor that is formed in one part of the world is gradually carried hundreds of miles away. Were it not for the various processes by which the vapor is condensed to the liquid or solid state again, the atmosphere would have long ago become completely saturated. But the limitation of evaporation chiefly to the under surface of the atmosphere where it rests on the oceans, coupled with the various means of condensation, generally prevents the occurrence of saturation through the great body of the atmosphere; and in continental interiors, far from the great oceans, as well as in the lofty atmosphere high above sea-level, the air is prevailingly dry.

177. Geographic and periodic variations of absolute humidity. The quantity of vapor in the lower air in the doldrums over the oceans is on the average greater than ~~elsewhere in~~ the world, because of the intensity of insolation, the presence of a water surface, and the quietness of the air. It is constantly near the value of saturation. In the trade wind belts on either side of the doldrums, the vapor is of somewhat less quantity; not so much on account of their lower temperature, as from their continual motion, whereby the moist lower currents are mixed with the next overlying and drier currents. In the horse latitudes, still lower measures of humidity are found, partly by reason of the lower temperature, partly because of the gentle downward settling of the air in these belts from high levels where the humidity is small. They are thus in strong contrast to the doldrums. In passing poleward through the prevalent westerly winds, the absolute humidity falls as the temperature decreases and in the polar regions its amount is very small compared with that of the torrid zone.

On the lands, the absolute humidity varies first with the prevailing temperatures, second with the distance from the oceans from which the prevailing winds blow, third with the degree of enclosure by mountains, and fourth with altitude above sea-level. The forested lowlands of the Amazon possess damp winds flowing in from the moist regions of the torrid Atlantic. The northwest coasts of Europe and North America have nearly as high an amount of vapor as their temperatures will allow in winter, when their winds come from the warmer currents of the oceans next westward. The interior of Europe and the lowlands of western Asia have progressively less and less humidity, because the vapor brought from the Atlantic has been in part abstracted on the way by condensation, and the rest has been mixed with dryer upper winds. Our far northwestern plains, whose winds come chiefly from the west, have low humidity, not only because they are far inland from the Pacific, but also because of the mountain barriers to windward; for mountains are very effective in abstracting vapor from the air. The greater part of our interior basin is a desert, in spite of its moderate distance from the ocean, because of the great height to which the Sierra Nevada rises

between it and the Pacific. The interior depression of Asia, enclosed on all sides by lofty mountains, is an arid waste because so great a share of the vapor in the winds that approach it has been left on the outer slopes of the ranges.

The quantity of vapor in the upper air is small; it rapidly diminishes as one ascends to greater and greater altitudes. Even in regions where the lower air is well supplied with vapor, the upper strata possess very little. This is partly due to the low temperature at great heights; but it depends also on the small ratio of the evaporating surface of the ocean to the volume of the air, and on the slowness with which vapor spreads by diffusion through the air. When the pressure of vapor at sea level is one inch, or one thirtieth of the total pressure of the atmosphere, it cannot be inferred that the vapor present at high levels exerts the same share of the pressure there experienced. If a perfect calm prevailed, such a condition might be approached, but numerous observations in the moving atmosphere on mountains and in balloons have shown that the decrease of vapor pressure upwards is much faster than the value of vapor pressure at sea level would indicate. The density of the vapor in the lower air is maintained, not only by the pressure of the vapor above it, but also by the resistance that the atmosphere opposes to the free upward diffusion of vapor from the lower strata.

The amount of vapor in the air is generally greater in the warmer season than in the colder. This is particularly the case in those regions where the warm season has inflowing winds from the sea, as in India. There the summer monsoon is a warm, moist, vapor-bearing wind, while the winter monsoon is cooler and comparatively free from vapor.

The average diurnal variations of the amount of vapor are small. The amount is somewhat greater by day than by night in regions where evaporation is hastened under sunshine; but such changes are greatly exceeded by the accompanying unperiodic shifts of the winds, by which the air is brought over the observer from a new source. Thus our southerly warm winds bring a large amount of vapor from the Gulf of Mexico and from the warm ocean waters near our southern Atlantic states; while our cold northwesterly winds bring little vapor from the continental interior, as will be further explained in the chapter on Weather.

178. Geographic and periodic variations of relative humidity. The variations of relative humidity are often unlike those of absolute humidity. The warm doldrums have a high relative (83 per cent or more) and a high absolute humidity; hence the air is sultry and oppressive. The trade winds over the ocean have a lower relative humidity (77 per cent) than the doldrums: on land, if they blow over a rising surface, they are moist; but if they pass over lowlands, they are dry and the region is reduced to a desert. The quiet air of the horse latitudes is somewhat drier than that of the trades. The

westerly winds, with much lower absolute humidity than the trade winds, have a high relative humidity over the oceans (90 per cent or more), and in their stormy areas they are saturated with vapor. A low relative as well as a low absolute humidity has often been recorded in the north polar regions. The upper air is relatively dry, as has been explained above.

The periodic variations of relative humidity are as a rule the reverse of those of absolute humidity. In the warmer season, the capacity for vapor generally increases in a higher degree than the amount of vapor. Thus continental interiors are as a rule drier in summer and by day than in winter and by night. If observations are taken in rapid succession through a day, the considerable variations that may be detected are often to be ascribed to convectional currents, by which masses of air from upper and lower levels are alternately carried past the observer.

179. Effect of water vapor on the general circulation of the atmosphere. The prevailing excess of water vapor in the lower equatorial atmosphere has a small effect in aiding the circulation of the general winds. Recalling the greater elasticity of vapor than of air, it follows that where vapor is in excess the isobaric surfaces will be held a little further apart than where it is deficient; recalling further that the absolute humidity over the great oceanic surface of the world increases rapidly with the temperature, it follows that the divergence of the isobaric surfaces already explained as a consequence of high equatorial temperatures (Section 111) will be a little further increased in consequence of the high equatorial value of the absolute humidity; and that a similar assistance will be given to the continental winds of winter when the air over the oceans is moister as well as warmer than over the lands. This effect is, however, insignificant when compared with that of the equatorial and polar or the oceanic and continental contrasts of temperature. It may be confidently asserted that if the earth had no oceans, and hence no currents by which the contrast between torrid and frigid temperatures is reduced, the poleward temperature gradients would be much stronger than we find them; and the gain in the velocity of the terrestrial circulation thus produced would much more than compensate for the loss following the withdrawal of the aid now given by the mere presence of water vapor.

It is important to notice that as far as water vapor acts on the circulation of the winds, the effect varies with the absolute and not with the relative humidity of the air. In spite of the high relative humidity of the damp and chilly atmosphere in high southern latitudes and the apparent great amount of vapor there present, the actual amount of vapor is much less than in the clear air of the warm trade winds. It is therefore inadmissible to ascribe the low Antarctic pressures to the presence of water vapor, as some have done.

CHAPTER IX.

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32°, the latter at lower temperatures. As vapor is thus withdrawn at the bottom of the atmosphere, an additional supply is furnished by downward diffusion from above, and condensation is maintained continuously on the cooling surface. In this it is like the formation of water drops upon the cold surface of an ice-pitcher in a warm room: the warm air of the room corresponds to the great body of little-cooled air above the earth's surface; the cold surface of the ice-pitcher corresponds to the cooling surface of the ground at night; and the drops of water upon the pitcher represent the part of the dew that is condensed from the air upon the ground.

182. Cooling retarded by the liberation of latent heat. Whenever water vapor is condensed, there is as much energy set free as was expended in the production of the vapor. This is called the liberation of latent heat. The greater intensity of a scald from steam at 212° than from water at 212° is thus explained. It has been stated in Section 167 that the latent heat required for evaporation may be supplied directly by the energy of insolation without taking the intermediate form of heat. It is equally true that the latent heat liberated in condensation may at once take the form of radiant energy and not appear as heat at all. In the example of the preceding section, the liberated latent heat does not raise the temperature of the body on which condensation takes place; it merely supplies a certain share of the radiant energy that is emitted from the surface of the body, and thus diminishes the rate at which radiation is supplied from the heat of the body itself. Hence the decrease of temperature both of the surface on which the air rests and of the lower air also is slower after condensation begins than before; and it is partly for this reason that moist regions have a small diurnal range of temperature; their greater cloudiness or haziness, and the retardation of warming by evaporation in the day-time and of cooling by condensation at night all conspire to this end. Arid regions, on the other hand, have extreme diurnal ranges of temperature. Southeastern California and the adjacent part of Arizona have an average summer diurnal temperature range of 40° or 45°; the greatest known in the world. The retardation of cooling by the latent heat of condensation is an extremely important matter in meteorology and will be frequently encountered in later pages.

183. Dew. The formation of dew has given rise to much discussion, but since the early years of this century, the experiments of Wells have generally been accepted as giving a satisfactory explanation of its origin, substantially as stated in the preceding paragraphs. Recent experiments by Aitken and others show, however, that only part of the dew and frost formed on the ground comes from the air; the rest comes from the ground or from plants. During the day-time, under sunshine and in the presence of an active wind,

the surface of the ground is dried; water rises from the subsoil by capillarity to supply more vapor to the thirsty air. The sap passing from the leaves of plants is also easily disposed of then, because it is exuded in the form of extremely minute drops, and because it is generally well exposed for evaporation to the surrounding air. But at night, drops of exuded water may collect on the leaves of grass and low plants, where it is unable to evaporate in the cold nocturnal air; and water rising to the surface of the bare ground may remain there, instead of passing off as vapor. If the ground is covered with grass, the blades of grass become colder than the ground beneath them; the vapor rising from the ground will then be in good part condensed on the cold grass.

There is a large variation in the proportion of dew supplied from its several sources at different times and places. In damp countries, the share coming from the ground is large; in dry regions, where the dew formed at night may be a large part of the water on which the growth of plants depends, the greater part of it presumably comes by condensation from the air. Few careful observations have yet been made on this subject.

184. Frost is formed in the same manner as dew, but at temperatures below the freezing point. The frost that forms just below the surface of the ground, raising the surface soil by the growth of its crystals, is probably derived for the most part from the subsoil: that which is deposited on loose leaves and sticks lying on the ground may also be supplied in good part from the ground; the share that then comes from the air is not yet determined.

Little attention has been given to the measurement of the diurnal or annual amount of dew and frost, on account of the difficulty of the problem. It has been estimated that the dew of a single night equals 0.02 inch of rainfall (Sect. 283); and that the total annual amount of dew precipitated in Great Britain would measure an inch and a half in depth. Further study should be given to this subject.

185. Conditions for the formation of dew or frost. In dry regions, even when the range of temperature is great and the nights are much cooler than the days, it may be that the dew-point is not reached at night and no dew is formed. This is commonly the case in deserts; but in regions of moderate aridity, where the rainfall is of small quantity and the days are warm and dry, the nights may often be cool enough to form dew or frost, as on our western plains, and on the pampas of the Argentine Republic. In well-watered regions, such as the eastern United States, the formation of a moderate amount of dew or frost in favorable situations is common on all clear and quiet nights, and it is accompanied by inversions of temperature of greater or less strength. The damp air of the torrid zone yields abundant dew on many equatorial lands

in the so-called "dry season," when the sky is prevailingly clear. The weather exerts a strong control on the formation of dew. The cooling of the ground and of the lower air will be greatest on clear nights, when there is no haze or cloud to diminish the effective radiation to outer space. The cooling will be greater on calm than on windy nights, for when the lower stratum of air lies quiet it is cooled more and more as the night goes on; while when the wind blows, a stratum that has been somewhat cooled by contact with the cooling ground is carried away, and replaced by another stratum from above that is less cooled. The special case of mountain frost-work in windy weather is described in Section 193. The nocturnal cooling of the ground has already been described as greatest upon convex surfaces, from which radiation goes on in all directions; and less from concave surfaces or valleys, above which there is a less surface of open sky. We might at first infer from this that hills would receive more dew than valleys; but it should be recalled that, in consequence of the surface cooling, a descending drainage is established from the hillsides to the valleys, so that the cold air of the hilltops is drained away and continually replaced by uncooled air. Hilltops therefore receive comparatively little dew. On the other hand, the air that creeps slowly down the slopes into the valleys is continually cooled by conduction to the ground as it moves on. It is true that at the same time an increase of pressure upon it tends to raise its temperature by compression; but as the descent of the air is generally rather slow, this cause of increase of temperature is less effective than the cooling by conduction to the cold ground during the gradual descent. Yet again, if the downward nocturnal drainage be concentrated in a narrow valley, leading from a large upland surface to an open lowland, the active mountain breeze blowing out of the valley may descend so rapidly as to reach the lowland less cooled than the air which crept more slowly down the adjacent mountain slopes; in such a case, the lowland opposite the mouth of the valley might be freer from frosts than the general lowland surface thereabouts. These varied examples afford good illustrations of the many processes at work to determine the formation of so simple a matter as dew.

186. Prediction of frost. The occurrence of frost in the late spring or early fall is injurious to many crops, and it is often highly important that warning should be given in the afternoon of the probable formation of frost at night. The general method of foretelling by means of weather-maps the occurrence of clear and quiet nights, when frost is likely to be formed, will be explained in Chapter XIII; but mention may be made here of a simple method of prediction applicable by any farmer. If the afternoon shows a decreasing cloudiness and a weakening wind, the temperature of the dew-point gives a ready means of inferring the minimum temperature of the night: for when the dew-point is reached, further nocturnal cooling is retarded, and therefore if

condensation begins at temperatures above 40°, it is seldom that the minimum temperature of the night will fall to freezing; but no definite rule can be given in this case. Local experience is needed to determine the "safety limit" of the dew-point and its relation to the season and the weather. Special study might be profitably turned in this direction.

187. Protection from frost. When the occurrence of frost appears likely, it is often possible to protect plants from injury by building a smoky fire on the windward side of a field, so that a dense stratum of smoke may drift slowly over the surface. Radiation is then transferred in great part from the ground and the plants to the smoky stratum, and the injurious fall of temperature at the level of the ground is effectively retarded. This method is often practised to protect tobacco in meadows and cranberries in low boggy ground, where the air is more quiet than on hills or slopes. On higher ground such protection is less often needed; for the more active movement of the air over hills, coupled with the conditions of nocturnal inversions of temperature and the nocturnal drainage of the air on sloping ground, generally serves to maintain the minimum temperature in quiet weather on hillslopes several degrees above that experienced in the neighboring valleys and lowlands. Peach orchards in the more northern states should be for this reason generally planted on rising ground: many examples could be given in which the blossoming in an orchard on higher ground escaped freezing, while another near by on lower ground had all its trees blighted. The limitation of low temperatures to the lower layers of the air is sometimes so marked that the upper branches of small trees or shrubs escape a frost that injures the lower branches.

188. Valley and lowland fogs. Valley bottoms are, in spite of the diminished nocturnal cooling of their soil, as a rule much damper at night than the adjacent hilltops; evaporation may continue from the little-cooled surface of their streams and ponds; they may not only receive a plentiful deposit of dew on the ground, but their lowest part may be covered with a layer of mist or fog. Common experience gives many examples of this; when driving late at night over a hilly road, the change of both temperature and dampness from hill to hollow is perfectly apparent. In the torrid zone, where the diurnal changes of temperature are peculiarly regular, the occurrence of nocturnal valley fogs is a characteristic feature of the climate. In the winter of temperate latitudes, when the long nights are interrupted only by short days of weak sunshine, the valleys of mountainous regions may be filled to a considerable depth with a cold damp fog; in spells of quiet and cold winter weather, broad lowlands are sometimes covered by fog sheets of this kind, extending over thousands of square miles, attaining a thickness of a

thousand or more feet, and remaining for even a week at a time; producing extremely raw and disagreeable weather beneath, while the sun shines above through air of extraordinary clearness (Sect. 249).

It sometimes happens that fog prevails even though the air is at a temperature several degrees above its dew-point. This is known as a "dry fog." It has been plausibly suggested that the fog particles of such a time may have an oily coating, derived from the combustion of coal and wood in cities, by which evaporation is retarded; but it is not yet proved that dry fogs occur only under conditions where the products of combustion are relatively plentiful. Like the occurrence of supersaturated cloudless air and of clouds formed of water particles, although the air in which they float may be several degrees below the freezing-point, as has sometimes been reported by balloonists, dry fog still awaits a full explanation.

All forms of fog may be classed between the dew and frost that are condensed at the bottom of the atmosphere and the overhanging clouds, to the consideration of which we now proceed.

189. Dependence of cloud condensation on "dust." Recent experiments by Aitken have given good reason to think that the formation of clouds in the open air is greatly aided by the presence of fine particles of solid or liquid matter, or, as it is commonly expressed, by the presence of "dust particles." If all suspended particles are removed from a volume of air, its temperature may be reduced several degrees below the dew-point before condensation begins, and the air is then said to be supersaturated. This is not of common occurrence in natural processes. In the lower air suspended particles are well known to exist in countless numbers. In the upper air at great heights, the particles that may be present are not properly named by the word dust; yet the best explanation of the blue color of the sky (Sect. 65) depends upon the presence there, as well as in the lower air, of matter not in the gaseous state, but in such excessively fine division as to remain suspended in the air for indefinite periods. If condensation of vapor into clouds requires the presence in all cases of solid or liquid nuclei, the nuclei must be of extreme fineness: for if rain-water is collected in a clear vessel, no perceptible sediment will settle from it, unless it is caught over some dusty or smoky region; yet every rain-drop has been formed by the union of countless minute cloud particles, every one of which, according to this theory, must have had a nucleus when its condensation began. While the laboratory experiments seem to leave no doubt upon this question, the natural occurrence of cloud and rain does not give it unqualified support: the hypothetical nuclei are not found by direct observation.

190. Size and constitution of cloud particles. Clouds formed at temperatures above 32° consist of minute spherical drops of water, $\frac{1}{4000}$ to $\frac{1}{1000}$ of

an inch or more in diameter. There is no sufficient reason for accepting the old belief that cloud particles are hollow vesicles. Clouds formed at temperatures below 32° consist of minute ice spicules, which may increase in size and become snow-flakes. The low temperatures at which such clouds occur prevail in the upper regions of the atmosphere all over the world, and at lower levels in the winter season or near the poles. Cloud particles are ordinarily so minute that they fall very slowly through even so light a medium as air, and a very faint ascending current is sufficient to bear them upward. When their size increases by continued condensation, they may become large enough to fall, slowly at first, faster afterwards; and thus rain or snow is produced. The association of rain or snow with storms of different kinds will be considered in the two following chapters, after which a later chapter will consider the occurrence and distribution of rainfall over the world.

191. Color of clouds. The numerous particles of which clouds consist generally reflect so large a share of the rays incident on them that they are of the same color as the light by which they are illuminated: they are therefore white in sunlight and gray in shadow during the day-time, or tinged with bright colors at sunset or sunrise. The brilliant light on the edge of a cloud that hides the sun is caused by the diffraction of rays on the marginal particles. At the time of formation or disappearance, light fleecy clouds often take a purplish tint when near the horizon and far from the sun. Distant massive clouds near the horizon may have a yellowish or even a ruddy tint at noon-time, on account of the selective scattering of the blue rays from the light that is reflected from them.

192. Halos and coronas. The icy nature of lofty clouds is known not only from observations on mountains and in balloons, but also by the action of the clouds on sunlight. A thin veil of lofty cloud (cirro-stratus) over the sky frequently produces a ring or halo of light around a less illuminated circular space of 21° radius, with the sun or moon in the center. The halo is colored when well defined, and then has red on the inside and blue on the outside. This may all be so well explained by reflection and refraction of light in ice crystals that there can be no doubt of the icy structure of such clouds. In the polar regions minute ice crystals are often scattered through the lower air. The halos then formed are of remarkable brilliancy and complicated form. Many dense and massive clouds must have a low temperature in their upper parts, and there they also must be of ice.

Coronas are concentric colored rings, ordinarily of small diameter, with the red on the outside, formed around the sun or moon in clouds of moderate thickness. These rings are produced by the diffraction and interference of

vesicular

light on fine particles of water or ice. The smaller solar coronas are generally lost in the blinding glare of light around the sun. Incomplete arcs of delicate coronal colors are often seen in the thinner margins of heavier clouds at much greater angular distances from the center of light.

193. Cloudy condensation in winds over cold surfaces. An effective means of cooling air to and below the dew-point is seen when our damp southerly winds of winter blow over a snow-clad surface. The snow is then rapidly thawed by the warmth gained from the air; at the same time the air is so rapidly cooled that it is not unusual for it to become foggy close to the ground. Fogs over the cold waters of the Newfoundland banks are formed in the same way. A small illustration of a similar process is sometimes seen in the formation of a cloud banner over a sharp mountain peak, when the air brushing past the summit of the mountain is cooled to a temperature below its dew-point. A standing patch of cloud is the result; its particles stream along with the wind beyond the peak, but before moving very far the moisture condensed by the cold of the mountain is re-evaporated by mixture with the adjacent air at the end of the cloud.

The cold rocks of mountains may at such times become heavily covered with frost, condensed upon them from the air. The frost grows to curious forms, building itself out to windward. On Mt. Washington the anemometer at the station of the Signal Service formerly maintained there was often so heavily covered with frost formed in this way as to be useless as a wind measure.

194. Clouds formed in poleward winds. Much larger illustrations of cloudy condensation occur in those winds which move from a lower to a higher latitude. Nearer the equator, where the air is exposed to stronger sunshine, its temperature is maintained at a higher degree; but as it advances poleward and the relation of insolation to radiation weakens, the temperature progressively falls, and the entire mass may become filled with a thick sheet of clouds, which lies like a cloak over thousands of miles of land or sea, and shelters the lower air from warming by day and from cooling at night. Condensation is aided when the poleward wind blows from an ocean over a winter continent; the air then cools by radiation downward to the ground as well as outward to space, and the cloud cloak becomes thicker. Such clouds are well known during southerly winds in the winter weather of the eastern United States. When their under surface is illuminated by oblique light, as at sunset or sunrise, one may see fleecy pendant masses, whose filaments gradually settle down and dissolve away in the lower warmer and drier air: when the clouds are very heavy, they may yield rain. If the cloud-making winds are stormy, fragmental clouds, or "scud," are formed

beneath the heavy cloud-sheet, leaning forward with the wind and rapidly changing their form. Northerly winds, whose temperature with us rises as they advance, and whose capacity for vapor therefore increases, are on the other hand characterized by a clear sky.

195. Condensation by the mechanical cooling of ascending currents. In all cases where a vertical or an obliquely ascending movement is given to a current of air, it is cooled by the expenditure of some of its energy in the work of expanding against the surrounding air as it rises, as has been explained in Section 49. This process, it must be remembered, is an immediate one, keeping close pace with the ascent: it is not like the process of cooling by conduction or radiation, which is slow. Mechanical cooling of ascending currents precisely accompanies expansion as the air rises to greater and greater altitudes. This is probably the most general process of cloud-making that occurs in the atmosphere. Attention was first called to it by Espy, a noted American meteorologist, about 1835; he was also the first to explain properly the prevailing clearness and dryness of descending currents, as well as the diurnal period of the wind and of cloudiness, now to be considered.

196. Convectional clouds of fair summer weather. Mechanical cooling of ascending currents is finely illustrated in the formation of day-time clouds — known as cumulus clouds — during the ordinary fair weather of summer time. The reader will recall from Section 54 the account of the local convectional disturbances of the lower air on warm summer days; and from Section 159, the application of this process to explain the diurnal variation in the velocity of the lower and the upper winds on land. We must now examine more carefully the formation of clouds whenever an ascending current rises so high as to reduce its temperature below the dew-point.

Let us consider the case of a summer morning, when the temperature of the lower air on the land close to sea level may be 70°, and its dew-point 67°, giving it a high relative humidity. As the sun shines through it and upon the ground below, the temperature of the lower air rises rapidly, and before nine or ten o'clock it may reach 75° or 80°. At the same time, evaporation from the dew-covered surface of the ground slightly increases the amount of vapor, but this increase is not nearly so rapid as the increase of capacity, and therefore, although the absolute humidity rises by a small amount as the morning advances, the relative humidity will fall distinctly. When the temperature reaches 80°, the dew-point may be 68°; the relative humidity will then be much lower than before. At this time, let it be supposed that the warmed lower air has become unstable, the vertical temperature gradient of the region being represented by the curve *BKN*, in Fig. 48. The lower air consequently ascends through the overlying cooler air; as it rises, it cools by

expansion at the relatively rapid rate of 1°.6 for every three hundred feet of vertical ascent, as explained in Section 49, and here illustrated by the line *BFK*. Section 197 will consider the rate of cooling when the dew-point is passed and latent heat is liberated by the condensation of some of the vapor.

When such a convectional process is well established, the ascent of the lower air is rapid and it reaches a considerable height. If the air, with temperature and humidity as above stated, ascends about half a mile, it will there be reduced almost to its dew-point; but it must be noticed that as the air ascends and expands, its dew-point falls slowly, because the vapor that is carried up will not so nearly saturate the increased volume of the expanded air as it did when the volume was smaller under greater pressure near the ground. The rate at which the dew-point is thus lowered by expansion is represented graphically by the dotted lines, *DE*, *de*, etc., Fig. 48; its numerical value being a third of a degree Fahrenheit for 300 feet of ascent (or 0.2° *C* for 100 meters). But the rate of cooling by expansion, *BJFK*, is more rapid than the falling of the dew-point, *DE*, and the former will soon overtake the latter in air of ordinary humidity: in the example here figured, saturation will be produced — not at a height *F*, where the temperature of the ascending air is reduced to the dew-point that it had before ascent began — but at a height *E*, or 2750 feet, and at a temperature of 65°.3; this being nearly 3° lower than the dew-point of the air before its ascent began. It may therefore be said that the height in feet at which condensation will begin in convectional currents equals the complement of the dew-point divided by (1.6 — 0.3) × 300.

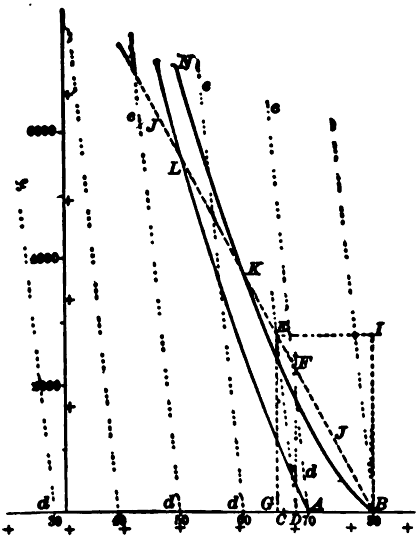

FIG. 48.

As the dew-point is reached and passed, saturation is followed by condensation, and the clouds of the morning appear. The dust that aids condensation in the open air is always present in ascending currents, unless the condition of the atmosphere is very exceptional; and hence there is as a rule no hesitation in the beginning of the process. It has, however, been supposed that if the ascending air should be extremely clean, the greater part of it might be cooled by expansion to temperatures distinctly below the dew-point, thus making it supersaturated, until at last a forced condensation takes place and produces a rapid and abundant clouding. But under ordinary circumstances

it must be admitted that cloudiness begins immediately after the dew-point is reached. The clouds are ragged at first; as they grow they gain a heaped-up form, and are therefore called cumulus clouds. They nearly always lean forward and curl over on reaching the faster-moving upper currents. As they absorb insolation that would otherwise pass down to the ground, their cooling is retarded and their bouyancy increased; this being an important aid to the ascent of all cloudy currents in the day-time. As far as the ascending current rises above the height at which condensation began, the cloud continues to grow: the further the cooling continues, the more vapor is condensed, the ascending air being always held at its dew-point. Attentive observers may easily detect the inflow at the base of these clouds, where misty wisps thicken as they rise and coalesce with the main mass; the ascent aloft where the convex cloud heads grow outward with visible expansion; and the melting away of the cloud as its foremost parts roll over and slowly descend.

The vertical thickness of the cloud will depend on the dampness of the air and the activity of its convectional ascent. In very dry air, such as occurs over deserts, the convection may be active in the lofty ascent of dusty whirlwinds, yet no clouds appear because the cooling by expansion is not enough to reduce the dry surface air to a temperature low enough to cause condensation. On the other hand, over the ocean where the air is moist, the height of incipient condensation may be less than a thousand feet. The flat under surfaces of these clouds, at the altitude at which condensation begins, are of about the same height all across the sky, and constitute as marked a feature of fair summer weather as the growing convex summits of the clouds. The height to which the cloud will grow above the level of first condensation depends on the additional height to which the ascending current rises, and this depends in turn on the contrast of temperature between the lower air and the overlying strata through which it rises. If the lower air becomes very warm, as at noon-day in summer, the clouds may grow to a truly mountainous height; while in the early morning the convectional ascent is so moderate as to produce only small patches of cloud in the clear sky; and in winter, when the air is generally stable, clouds of this kind are uncommon.

On almost any fair summer day the convectional process of cloud-making may be watched from its beginning in the morning till it ceases about sunset. Small curling clouds and fresh breezes indicate the establishment of active convection at an early hour; the clouds increase in size through the morning, and if the ground has been moistened by a rain the night before, the adjacent clouds may grow to so great a size as almost to coalesce and overcast the noon-time sky: but the clear blue of the upper air may still be seen in the intervals between the clouds. With their growth, the morning breezes freshen and become a brisk wind in the afternoon; hence the common name of "wind clouds," often given to these forms. The strong sunshine passing through

spaces between the clouds lights up the dusty afternoon air with slanting beams of light, apparently converging to the sun, but really parallel. If the observer ascend a lofty mountain-slope on such a day, he may reach the level of condensation at which the bases of the clouds all rest.

With the shading of the ground by the cloudy cover, and with the descent of the sun in the afternoon, the cause of the convectional action is weakened, and after four or five o'clock it has generally stopped. As the ascending motion on which the growth of the clouds depends is lessened and as the weight of the water particles of which the clouds consist depresses the ascending currents, the clouds settle down and dissolve away. Thus the sky of late afternoon or evening is left as clear as it was in the morning. The regularity with which this process is repeated in fair summer weather, especially in the fair weather of the torrid zone, leaves no doubt of its dependence on diurnal sunshine and convection.

197. Decreased rate of adiabatic change of temperature in cloudy air. We must return now to examine the effect of the liberation of latent heat in ascending currents of air when cloud-making has begun; and of the reverse process in descending cloudy currents.

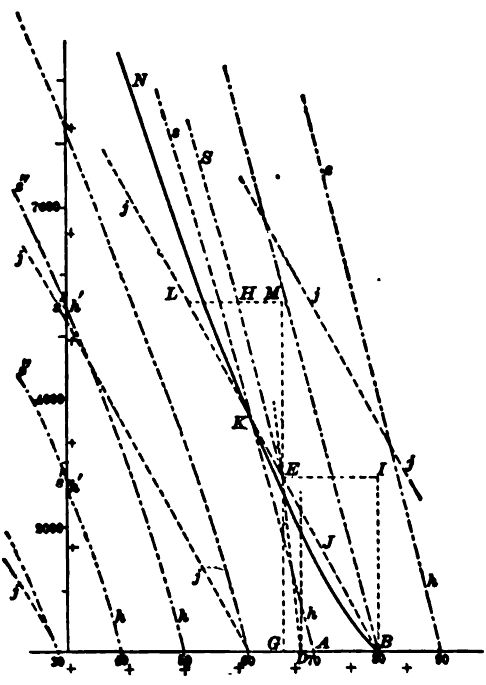

FIG. 49.

The retardation of nocturnal cooling when latent heat is liberated in the formation of dew or frost has been explained in Section 182. In that case the liberated energy passed away as terrestrial radiation into space, without doing any work on the earth. In the problem now before us, the liberated latent heat of an ascending, expanding current is applied to aiding in the work of pushing away the surrounding air; the heat of the ascending current is therefore drawn upon more slowly to do its share in this work, and hence the fall of temperature in the ascending air is retarded. This process is of wide application, and must be carefully considered.

In Fig. 49, let the altitude and temperature of incipient condensation in a mass of air that has risen from sea-level be indicated by *E*. In this case, we

may say that the work of expansion, represented by *IE*, has all been performed by the only available energy; namely, the heat of the rising air, and that the air has for this reason been cooled from the temperature *B* to *G*. If the air should now ascend through the additional height, *EM*, the corresponding work of expansion may be represented by *ML*. There are now two available sources of energy that may be called on to do this work; the sensible heat of the moist air, and the latent heat liberated from the vapor that is condensed in consequence of the cooling by the loss of the sensible heat. It is found by calculation from the known values of the capacity of air for vapor and of the latent heat of vapor, that a part, *MH*, of the total work, *ML*, requires a loss of sensible heat that will cause the condensation of an amount of vapor whose liberated latent heat will just perform the remainder of the work, *HL*. Hence when condensation begins, the sensible cooling of the ascending current will no longer be at the rapid rate, *EL*, but at the retarded rate, *EHS*.

The retarded rate of adiabatic cooling has been determined for various temperatures and pressures, as given in the following table, or as represented in Fig. 49 by the broken lines, *hs*. The greatest retardation is found at high temperatures and heavy pressures; that is, under conditions where the decrease of capacity with decrease of temperature is most rapid. At very low temperatures, the retardation is insignificant.

Rate of Cooling of Cloudy Ascending Currents.

A. Decrease of Temperature in Fahrenheit Degrees for 300 Feet of Ascent.

Pressure.	10°	20°	30°	40°	50°	60°	70°	80°	90°
30″	1.2	1.1	1.0	1.0	0.9	0.8	0.7	0.6	0.5
26	1.2	1.0	1.0	0.9	0.8	0.7	0.7	0.6	0.5
22	1.1	1.0	1.0	0.9	0.8	0.7	0.6	0.5	—
18	1.0	1.0	0.9	0.8	0.7	0.6	—	—	—

B. Decrease of Temperature in Centigrade Degrees for 100 Meters Ascent.

Pressure.	−10°	−5°	0°	+5°	10°	15°	20°	25°	30°
760	0.74	0.68	0.64	0.58	0.53	0.48	0.43	0.40	0.37
700	.73	.66	.63	.57	.51	.46	.42	.38	[illegible]
[illegible]00	.70	.63	.60	.54	.48	.43	.40	.36	—
[illegible]0	.66	.60	.56	.50	.45	.40	.37	—	—
[illegible]00	.62	.55	.51	.46	.41	.37	—	—	—
[illegible]0	.56	.49	.46	.42	—	—	—	—	—
[illegible]0	.48	.41	.39	—	—	—	—	—	—

If a mass of cloudy air is descending, the problem is reversed and the rate of adiabatic warming by compression is retarded. The heat that would be given entirely to raising the temperature of the air if it were not cloudy is in this case devoted in part to supplying latent heat for evaporation of the cloud particles. After all the cloud is evaporated, the rise of temperature goes on at the usual rapid rate of 1°.6 for every 300 feet of descent. An important application of this principle will be found in Section 248.

198. Special adiabatic conditions at the freezing-point. A peculiar condition is found in a convectional current whose ascent is so high as to cool it to the freezing-point, as often happens in thunder-storms. The cloudy current of air is saturated with vapor and carries upwards a vast number of minute water particles. When the freezing temperature is reached no further cooling can take place until all the suspended water particles are frozen; and during this time, all the work of expansion is done by the latent heat liberated in the change of the water from the liquid to the solid state. Still more; in consequence of the expansion of the air during this time of no cooling, its capacity for vapor slightly increases, and some of the water particles or ice crystals will return to the vaporous state, deriving their necessary latent heat from that liberated in the freezing of some of their neighbors. Hence for a brief space, ascent may be accomplished without decrease of temperature, as shown by the lines, $h'h''$, $s's''$, Fig. 49. No definite statement can be made concerning the height of ascent without cooling at this critical stage of the process, because the amount of water present varies with the size of the cloud and with the velocity of the ascending current; but in some cases it may amount to as much as fifty or a hundred feet. It is therefore at most a trifling matter, and only deserves mention from its theoretical interest.

199. Increased altitude of convectional ascent in cloudy currents. Recalling the explanations of Section 197, it is apparent from Fig. 49 that if no cloud were formed, the convectional ascent of the air there considered would cease at *K*, where the adiabatic line, *BJL*, intersects the vertical temperature gradient, *BKN*. But as cloud is formed after reaching the altitude *E*, the rate of cooling is then changed from *EKL* to *EHS* by the liberation of latent heat; and hence the ascending air will not be reduced to the temperature of the air through which it ascends until a much greater height than *K* is gained. The higher the temperature and the damper the air, the more effective is the aid thus given from the condensing vapor. The velocity of ascent is also increased after cloud-making begins; for on account of the retarded cooling of the cloudy air, its excess of temperature over that of the surrounding air is maintained at a greater value than it would be in an unclouded current. Application of this principle will be found abundantly in the chapters on storms.

It must be borne in mind that the simple adiabatic changes here considered are never precisely realized. Mixture, conduction and radiation all tend to equalize the temperatures of the ascending and surrounding air, and thus to diminish the altitude attainable; and the cloud particles act as a burden which holds down the ascending current below the height it might gain without them. On the other hand, the cloud particles absorb by day, when convection generally occurs, much insolation that would otherwise pass on to the earth, and this aids the retardation of cooling in the ascending current; and the momentum of the ascent tends to carry the current above the normal height of equilibrium, especially if the ascent be rapid. Even if all these disturbing influences could be allowed for, the value of the vertical temperature gradient could seldom be well ascertained, and hence the point of intersection of the two critical lines must be ill determined. The curves of Fig. 49 and of other figures of the same kind must be taken only to indicate a rough solution of the convectional problem; but if thus understood, they will be found of much assistance in gaining a clear idea of important meteorological processes. The problem of the formation of clouds may now be resumed with a fuller understanding.

200. Convectional clouds over islands and mountains. Mountainous islands are often clothed with diurnal clouds while the surrounding air is comparatively clear. This is because the combined action of the inflowing

FIG. 50.

sea breeze and the ascending valley breeze carries the damp air from over the ocean up to a height at which it becomes cloudy. A cloud-ring of this origin has been described over the island of Hawaii in the tropical Pacific ocean (Sect. 256).

Either of the two processes of cloud-making here in operation may produce clouds when acting alone. Arms of the land, like Cape Cod, may be marked out in the summer sky by the growth of floating cumulus clouds in quiet weather.. At the same time, the sky over the mainland to the northwest is heavily charged with large clouds, while over the sea the sky is clear, except that in quiet weather isolated clouds grow over the sandy island of Nantucket, Fig. 50, the island itself not being visible from the mainland. The sand-bars enclosing Pamlico and Albemarle sounds should determine the development

of clouds in calm summer weather in the same way. On the other hand, inland lakes may produce clear sky, when the surrounding land is cloud-covered. Inland mountain-ranges are generally obscured in the afternoons by the formation of clouds around their summits, often growing to the size of thunder storms. Mountain climbers, knowing this, make their ascents as early as possible to gain a more extended view.

201. Varied form of convectional clouds. The rate of ascent of the convectional cloud-forming current is often so slow and the burden of cloud particles is so heavy a load for it to bear up that the summit of the cloud fails to reach the altitude where its temperature would be reduced to that of the air about it, and where it might then spread out horizontally, after the fashion of dust whirlwinds. In such cases, the cloud topples over and dissolves away. Where the activity of ascent is moderate, and the horizontal dimensions of the cloud exceed the vertical, it is called strato-cumulus, Fig. 51. The top of a more active cloud mass sometimes spreads out laterally at a moderate altitude, forming a high, flat cloud with definite, sharp-cut margin. It is probable that this form is best developed when the altitude to which the cloud current rises is determined not only by its own cooling but also by reaching a relatively warm upper current, into which it cannot rise, and by whose more rapid forward motion the top of the cumulus cloud is brushed out into a horizontal sheet. Sheet clouds of this origin often float away for many miles, gradually breaking into alto-cumulus masses and slowly dissolving, but remaining visible long after their original cumulus source has disappeared. The flat layer clouds of sunset are often the remains of outspreading clouds formed in this way.

FIG. 51.

If the cumulus cloud rises into damp upper air, a thin sheet of cirro-stratus cloud may be formed over the rising summit of the cumulus mass, soon to be broken through as the cumulus rises higher.

If the convectional ascent be excessive, as in thunder-storms (Sect. 254), the cloud mass may attain an extraordinary volume and altitude, being then called cumulo-nimbus. Some such clouds have been charted over a length of two or three hundred miles, while their summit height has been determined at six or eight miles. A broad outflow of soft fleecy cirro-stratus cloud floats away from the top; it is generally unsymmetrical, reaching further forward with the upper wind in the direction of the advance of the storm. When fibrous at the edges, slowly curling and eddying, it is called cirrus. The under-

surface of the cirro-stratus cloud is sometimes festooned by the settling down of misty layers from its under surface into the clear air beneath. The lower portion of such a great thunder-storm cloud consists of water-drops; but the upper portion may be of snow even in summer, as is proved by the snow falling from thunder-storms on lofty mountains, while the valleys receive only rain. Examples of much larger cloud-masses in connection with tropical convectional storms will be given in Section 218; they need mention in this connection because of the great development of long, feathery, cirrus clouds that radiate from the lofty storm-cloud mass for many miles around, forming a stratiform shield at heights of five or more miles. They are matted together near the stormy area, and are there called cirro-stratus; further away they are more fibrous and feathery, and are known as cirrus. Their form changes very slowly. They appear to be the final product of the cooling by ascent, perhaps aided by mixture with the cold lofty air, as they are formed where the central ascending component of the motion gradually turns outward to a far-reaching, nearly horizontal movement. The small change observed in such clouds from day to night implies that warming by the diurnal absorption of insolation and cooling by nocturnal radiation has but a moderate effect in producing or transforming them; but the stratiform portion of these lofty sheets sometimes breaks up into patches, as if prompted to local convectional movement by arrested insolation; the little fleecy clouds thus formed being called cirro-cumulus (see also Sect. 203).

202. Clouds in forced ascending currents. The ascent by which clouds are formed need not be of spontaneous convectional origin, as in the cases just considered. Any process by which the air is raised to levels of less pressure will serve, if it continues far enough. Thus when the damp trade winds blow against a mountain slope, and rise to pass over it, they soon become cloudy. If the winds are strong and the mountains high, the clouds may grow so large as to give forth rain.

Promontories facing a windward sea are for the same reason frequently cloudy. Table Mountain at the Cape of Good Hope is spread over by a sheet of cloud, known as the "table-cloth," when damp winds blow over it; the cloud grows as the wind ascends on the windward side, and dissolves away as it descends to leeward. Clouds of this kind sometimes do not touch the mountain over which they are produced; and from this it may be inferred that only certain ones of the arching atmospheric strata are damp enough to pass the dew-point as they rise. It sometimes happens that a comparatively dry wind passes over a mountain crest and draws up damp currents on the leeward slope, which then become cloudy.

In the blustering winds of March, when local convection is becoming active with the rapid warming of the ground and the lower air while the

upper air is still cold, the rolling of the lower currents may often carry a convectional updraft for a short time to a greater height than it could reach in a still atmosphere; thus forming the ragged clouds of that windy season. Many of the tangled cirrus clouds (Fig. 52), floating at great heights and changing their form slowly, may perhaps be ascribed to a driven ascent in conflicting currents (see also Sect. 208). Indeed, reason will be given in a later chapter for thinking that many of the great masses of heavy clouds with long, outspreading cirrus plumes, measuring one or two hundred miles in length,

FIG. 52.

that form over the stormy areas of the prevailing westerly winds of the temperate zones, particularly in winter time, may be in great part examples of condensation in a driven whirling ascent on a large scale (Sect. 237).

203. Clouds formed in atmospheric waves. When adjacent air currents move with different velocities or in different directions, their surface of contact may be thrown into a series of slowly oscillating waves of considerable horizontal length from node to node, but of moderate vertical amplitude of oscillation. The waves of the sea surface are produced by the difference in velocity of wind and water; the ripples on the surface of sand dunes and of snow drifts, as well as of sand bars under water, are of similar origin; the violent agitation of the sea, known as "rips," where tidal or other currents conflict, are of the same kind. Waves in the atmosphere are generally of very slow movement. Their existence may be recognized in the undulating form of the under surface of broad cloud sheets. The regular spacing of clouds between equal intervals of clear sky also depends on some undulatory movement of atmospheric strata. Indeed, atmospheric wave motion is

probably much more common than we imagine; for we notice it only when it becomes conspicuous by means of clouds produced or shaped by it.

The formation of clouds in waves depends on the variation of density, and hence of temperature, produced by the vertical component of their motion. The vertical oscillation, by which a portion of air is carried from the trough to the crest of the wave, may in certain cases cause sufficient cooling to produce cloudiness. If so, the cloud should increase on the front of the wave and fade away on the rear. Condensation once begun in this way may give rise to further cloud growth by arresting sunshine; but this whole process calls for detailed observation before it can be considered well understood.

The forced passage of the wind over an obstacle, such as a mountain-ridge, sometimes throws the current into standing waves for some distance to leeward. In northwestern England, when a damp wind blows over the Cross Fell range from the east, a cloud is formed over the ridge by the forced ascent of the air, and a second cloud, known as the Helm Bar, is formed in the ascending part of a standing wave of wind, a short distance to leeward. Similar clouds may be expected in the damper mountainous parts of our western country, although they have not yet been recognized.

Much larger examples of clouds formed in standing waves have been described over and to leeward or northwest of the island of Ascension, where they float in the trade wind for fifty or more miles in the ocean. It is suggested that even lofty cirrus clouds may be formed by the ascent and undulation of high currents where their even flow is disturbed by the arching of the lower currents over islands or mountains; but, like many other suggestions in this chapter, much more observation is needed for the full confirmation of this process.

204. Clouds do not always float with the air currents. It is commonly assumed that the movement of clouds gives a direct indication of the movement of the air in which they are suspended; but a number of examples described above show that this is not necessarily the case. The apparently fixed, level base of a cumulus cloud is really the site of a comparatively active ascending current. The stationary cloud-banners that sometimes stream out to leeward of mountain peaks merely indicate the space within which the moving air is reduced to a temperature below its dew-point; the air flows rapidly along, bearing the individual cloud particles with it, but the cloud stands still. The same is true of clouds formed in fixed waves, determined by irregularities of the land. Finally, the movement of those high-level clouds that seem to be formed in the rippling wave crest of adjacent air currents must differ by a certain unknown amount from the motion of the currents that produce them. A later section (272) will give illustration of a cloud whose change of outline actually progresses *against* the movement of the wind by

which its particles are carried. These facts should be born in mind and allowed for as far as possible when observations of clouds are employed in the study of the movements of the atmosphere.

The striation of lofty fibrous clouds seldom indicates the direction of their movement with respect to the earth's surface. They frequently drift transversely to their length; and then their trend as well as their drift should be recorded. The twisted wisps often seen on the under surface of lofty clouds are generally due to the sinking of cloud particles from one current into another of different direction and velocity; the direction of the wisps then indicates the movement of the lower current with reference to the upper current; just as the smoke from a moving train is not parallel either to the railroad or to the wind, but closes the triangle of which the train and wind movements are two sides.

205. Condensation caused by conduction. It is possible that certain thin sheets of cloud may be formed in the warmer of two adjacent air-currents whose temperatures are distinctly different. Conduction causes the temperature of each current to approach that of its neighbor; and the cooling of the warmer current may make it cloudy. Like several other processes here considered, no definite value can be assigned to this one; but all possible processes should be borne in mind when undertaking this most difficult division of meteorological observation.

206. Condensation by the upward diffusion of vapor. The occurrence of diurnal convectional currents has been explained as depending on the overwarming of the lower air under sunshine. It is most marked on land surfaces and in summer, when the diurnal range of temperature is strong. At sea, where the range may be generally less than four degrees, the increase of temperature in the lower air does not appear to be sufficient to cause instability and convection; and yet in the trade belts and especially in the doldrums over the oceans, local cumulus clouds are regularly formed in the morning, and rise to great heights in the afternoon, generally causing rain. It may be suggested that these clouds are due in good part to an upward diffusion or expansion of water-vapor, formed in excess at the surface of the ocean; or at least that this process is of decided importance in combination with any convectional motion that may take place there. A slender form of the lofty cumulus clouds in the trade winds is said to be characteristic.

207. Condensation by radiation from the atmosphere. The atmosphere has already been described as a poor radiator; its temperature falls but little at night, because it cannot easily give up its heat by the emission of radiant energy. It is therefore somewhat uncertain whether the sheets of cloud,

sometimes observed in the early morning after a night that was clear till a late hour, can be ascribed to so ineffective a process as the cooling of the air by its own radiation. It may be, however, that certain layers of air are so nearly at their dew-point in the day-time that the little cooling of night makes them cloudy. If the process is once begun, its extension is easy; for the cloud particles themselves will cool by radiation, and the air near them will then cool by conduction and radiation to them; condensation once established may be rapidly extended. A possible cause for the moist condition of certain strata of the atmosphere is found in the outspreading of cumulo-stratus sheets from the top of cumulus clouds, as explained in Section 201. The cumulo-stratus sheet slowly dissolves away in the late afternoon, thereby dampening the layer of air about it. Thus the transfer of vapor from the lower air or even from the ground in the day-time may supply the vapor for the formation of high-level cloud sheets by radiation late in the succeeding night; but like several other suggestions regarding the origin of clouds, this is altogether problematic. It seems to be physically possible, but its value in actual occurrence is undetermined.

208. Condensation by mixture of air currents. If two masses of air, both saturated but of unlike temperatures, are thoroughly mixed, the temperature of the mixture will be below its dew-point, and a general clouding of moderate density will be the result. This process of condensation was suggested in the last century by Hutton, and was for many years regarded as the most effective means of producing clouds and rain. It is still often referred to, but it cannot now be regarded as so important as several of the processes already considered.

The present understanding of meteorological phenomena shows that Hutton's theory involves an uncommon process, and that it is of relatively little importance when it occurs, except as a subordinate aid to other processes. It is uncommon, because it seldom happens that two adjacent currents of unlike temperatures are both saturated. Even if this condition occurred, the intimate mixture of the two currents is not easily brought about. If mixture should take place, the resulting condensation would not supply clouds and rainfall in observed amounts, unless improbable temperatures, volumes and velocities are assumed for the intermixing currents. Little attention is therefore now given to this process alone as a means of producing so active a condensation as to cause rain. Coupled with other processes, it has some undetermined value. Smaller examples of its action may perhaps be seen in the formation of the tangled cirrus clouds mentioned in Section 202, although the prevalent dryness of the upper air is against such an explanation; or of the thin wave-like cloud films that are sometimes seen in our southerly winds when they flow over colder lower air; in this case the surface of contact is recognized by the wave-like form of the cloud film. In a larger way, mixture must aid in the

formation of the great masses of storm clouds from which most of our winter rain and snow falls; but it should be noticed that in such cases the various processes by which the mixing currents have been cooled to their dew-points continue in operation after mixture as well as before, and that the greater amount of condensation is to be expected from these continuous processes rather than from mixture, whose cloud-making ceases when the intermingling is once accomplished.

The process of mixture of different air masses is more often the cause of the dissolution of clouds than of their formation. When an ordinary morning cumulus cloud rolls forward as it rises and its margins mingle with the higher air that its ascending currents enter, the mixture of the saturated cloud-bearing streams of air with larger volumes of clear and relatively dry air alongside, ordinarily enables all the condensed vapor to dissolve again. The same process must be common in lofty cirrus clouds, whose minute and thinly-scattered ice crystals are frequently evaporated along the feathery margin of the cloud, where, according to the theory of cloud-making by mixture, the cloud should be densest.

It may be well to refer briefly to another inefficient process often mentioned as producing clouds. This is the ascent of warm, damp lower air into the "cold regions of the upper atmosphere." It is true that the upper air is cold, and that ascending currents become cloudy, but there is no reason to think that any large part of the cloud mass is due to cooling by conduction to or mixture with the cold and generally dry upper air; for if so, an ordinary cumulus would be only a hollow shell of cloud. The ascending current becomes cloudy by reason of its own mechanical cooling, as has been fully considered on earlier pages.

209. Haze. There are all gradations from a sky of perfect clearness to turbidity of varying degrees of density, known as haze. This is sometimes the product of forest fires, such as are frequent on our western mountains in dry summer weather, when the transparency of the air for hundreds of miles to leeward is lost for weeks together, and only the nearer hills remain in sight. In northern Europe, the smoke from the burning of extensive peat bogs sometimes dulls the sky over large regions. Haze is sometimes caused by the presence of very fine mineral dust, gathered from deserts and suspended or carried far away by the winds. The islands west of the Sahara are often thus afflicted. In our own country, the warm southerly winds of spring and summer are often hazy, with a glaring sky of pale blue; the haze is thought to be largely composed of water particles, but the cause of their condensation is not understood. True water vapor is entirely invisible. Fine water particles are more easily evaporated than larger ones, not only because they have small volume, but also by reason of the sharp curvature of their surface;

it being proved by experiment that evaporation may be continued from a convex surface after it ceases from a plane surface. The conditions of the formation and duration of haze therefore still need examination. When the haze pales the blue of the upper sky, and yet leaves distant objects visible through the lower air, it is sometimes called cirrus haze.

210. Conditions that favor clear sky. It is pertinent to introduce in this connection the opposite question of the causes of a clear or clearing sky. Probably the most effective cause is a scarcity of water vapor, such as characterizes interior desert regions, far removed from oceans and enclosed by lofty mountains. The skies of such regions are generally dusty, but not cloudy: they form the natural contrast to the prevailingly clean but cloudy oceanic skies. The higher strata of the atmosphere are never clouded; the quantity of vapor that can exist there is very small, and no effective cause of condensation seems to operate at great heights. Convectional action and rapid changes of temperature are, as a rule, limited to the lower atmosphere.

The occurrence of descending air currents is practically prohibitive of cloud making. It is true that the compression of a mass of moist air, whose temperature is maintained at a constant value, will soon cause condensation of vapor into mist; but the natural process of compression during descent is always accompanied by the generation of heat and a consequent rise of temperature, which effectually counteracts the tendency to condensation due to decrease of volume; except in those special cases where the descent becomes very slow, as on approaching the ground in winter, and where the heat gained by compression is lost by radiation and conduction to the cold surface of the earth, producing heavy fogs in cold weather (Sect. 249). The trade winds, or the northerly winds that follow our spells of cloudy and rainy weather (Sects. 294, 315), approach the equator and gain a higher temperature which effectually dissipates any clouds that they may at first have borne; and such winds are prevailingly clear, unless prompted to roll over by local convection or irregularity of path. The flow of a cool wind, bearing fog from an ocean upon a warm summer land, soon brings about a sufficient rise of temperature to evaporate the fog. Quiet air, or air without vertical components of motion, allows cloud particles to settle slowly to levels of higher temperature, and therefore soon becomes clear; as on calm summer evenings, when the clouds produced by convectional action during the hotter hours of the day sink down and fade away, leaving a clear, star-lit sky. Conduction between adjacent air currents, appealed to already to produce cloud sheets, may under favorable conditions cause them to disappear; for if a sheet of cloud, borne by a cool current, comes in contact with a warm and dry stratum of air, the warmth gained by the former from the latter may dissolve the cloud away and leave both currents clear. The mixture of two air masses, one of which is cloudy,

is quite as likely to produce clear air as to increase the cloudiness; the disappearance of clouds on the rear of our stormy areas when fair weather is approaching may be in good part due to this process.

211. Classification of clouds. The previous sections have described the forms of clouds produced in different ways. The descriptions have, however, involved certain hypothetical explanations, which may not in all cases be correct; these may serve as suggestions for deliberate investigation, but they are not serviceable as guides in recording the occurrence of clouds at the hours of regular observation. It is therefore advisable to classify clouds in some simple manner for convenient record.

The classification adopted by the Signal Service in this country, and still in use in the Weather Bureau (Sect. 318), is slightly modified from that of Howard, proposed in 1803. A somewhat different system was recommended by the International Meteorological Congress held at Munich, in 1891. The following table, prepared by Mr. H. H. Clayton, indicates the relations of the two systems: —

Weather Bureau.	International Congress.
Cirrus	Cirrus
Cirro-stratus	Cirro-stratus (alto-stratus)
Cirro-cumulus	Cirro-cumulus Alto-cumulus
Cumulus	Cumulus
Cumulo-stratus	Strato-cumulus Cumulo-nimbus
Nimbus	Nimbus
Stratus	Stratus

The following descriptions apply to the terms adopted by the International Congress. It is very desirable that uniformity the world over should be gained in terms of this kind. Comparison of observations is otherwise impossible.

Cirrus clouds consist of slender fibres, sometimes in long parallel lines, sometimes in feathery, curled, tangled or clotted arrangement. Their form changes slowly and their movement is apparently not so fast as that of clouds at lower levels; but as their altitude varies commonly between five and ten miles above sea-level, their actual velocities may be rapid, from 50 to 100 or 200 miles an hour. Cirrus clouds, as a rule, drift eastward (Sect. 147); but they occasionally advance slowly to the west in connection with storms.

Cirro-stratus clouds. True cirrus fibers are often associated with horizontal cloud layers at similar or slightly less altitudes, as if formed by the matting together of growing filaments. These layers are often extended in bands of considerable length, sometimes in parallel trains, straight or gently curved, and associated with true cirrus fibres; they may reach all across the sky and

winds at a rate of eight, ten, or twelve miles an hour. The barometer continually sinking, and at last falls rapidly; with this the roaring wind increases to full hurricane strength, the low scud clouds fly before its blasts, the lightning flashes, the rain descends in drenching torrents, cooling the sultry air. All the elements are in uproar; yet only a day or two before there may have been no sign of the coming storm, except the ominous heaving of the sea.

Before the law of storms was learned, many a ship was borne before such a hurricane, with all sails furled or blown away, helpless in the violence of the winds and waves; and when the vessel was at last about to founder, the wind has suddenly weakened to a calm in the eye of the storm; falling from its greatest violence to an almost perfect repose in fifteen minutes or less. The rain ceases, even the clouds may break away, showing the blue sky by day and the stars by night; but the waves still roll and toss, and in even more dreaded form than in their regular heaving before the hurricane; for in the eye of the storm they swing in from all sides, and pitch and heave tumultuously, forming irregular pits and peaks of water which strain a vessel violently, even to leaking and sinking. A few careful records made in the calm storm center while passing over a land station show a peculiar change in the temperature and humidity of the air. Underneath the surrounding heavy clouds, the air is somewhat cooled and held close to its dew-point by the rainfall; yet the air within the calm center has been found to be comparatively dry with a temperature unduly high; but it is not yet known if these features always prevail. The diameter of the calm space may be ten, twenty, or thirty miles, perhaps a tenth or a fifteenth of the diameter of the whole storm; and its duration in passing a given point may vary from half an hour to two hours: the barometer reading in the center may be even less than 27 inches.

As the hurricane on the further side of the central calm approaches the observer, its moaning can be heard in the distance, rising to a portentous roar as it comes near, and then breaking suddenly with as great fury as the hurricane which died away before, but its direction is now the reverse of that of the winds by which the calm was preceded. All the elements of the cyclone now re-appear; the blasts of the wind beat up the waves to their greatest height, the clouds hang low and heavy over the darkened sea, the rain falls again in torrents; and then as the storm gradually moves away, all these signs of its activity weaken. In the course of a day or two, the barometer rises nearly to its usual height, the wind dies down, the waves fall to a long low swell, the lower clouds recede, the lofty cirrus plumes retreat after them, and the sky is left in its accustomed clearness.

219. Law of storms. Until about 1830, there was little knowledge of the systematic courses followed by the winds in cyclones, and ships at sea were at

by alto-stratus. As this refers rather to the state of the weather than the kind of cloud, the term is not entirely satisfactory. An observer on a high hill, receiving rain from a dripping cloud not far overhead, would call it nimbus; while an observer on an adjacent lowland might call the same cloud by some other name. A heavy cloud sheet, hanging at a moderate height and threatening rain, would be called stratus by an observer where rain had not begun; and nimbus by another a few miles away where rain was already falling. It is often preferable to employ some indefinite term, as "overcast," when the cloud evenly covers the whole sky and its nature cannot be determined.

212. Altitude of clouds. Simultaneous observations of the angular altitude and azimuth or direction of clouds made by two observers communicating with each other by telephone from stations a mile or more apart serve to determine the height at which the clouds float, their dimensions, and the direction and velocity of their motion. Simultaneous photographs of clouds from different stations have been used in the same way. This style of observation might be taken up to advantage by observers who can provide themselves with suitable instruments for angular measurements, and can use a telephone connection between their stations.

The following series of altitudes (in meters) have been determined by recent measurements in Sweden, and at the Blue Hill Observatory near Boston; the comparative table being prepared by Mr. Clayton, observer at the last named station.[1]

Kind of Cloud.	Blue Hill, Massachusetts.					
	Summer Height.			Winter Height.		
	Mean.	Max.	Min.	Mean.	Max.	Min.
	Meters.	*Meters.*	*Meters.*	*Meters.*	*Meters.*	*Meters.*
Cirrus	9923	14930	5392	8051	11560	3764
High Cirro-stratus	8754	12134	5521	7846	8512	6823
Low Cirro-stratus	6481	12050	2290	2930	. . .	. . .
Cirro-cumulus	7606	10520	4772	6992	8570	4571
High Alto-cumulus	6406	8204	3110	. . .	. . .	. . .
Low Alto-cumulus	3168	7047	784	2884	. . .	. . .
Strato-cumulus	2003	3328	1100	. . .	. . .	. . .
"False Cirrus"	8242	12360	5392	. . .	. . .	. . .
Cumulo-nimbus (top)	. . .	. . .	. . .	. . .	. . .	. . .
Cumulo-nimbus (base)	1202	1590	884	1552	2058	1046
Cumulus (top)	2181	. . .	1455	. . .	. . .	. . .
Cumulus (base)	1473	3582	601	1381	2690	532
Nimbus	712	1720	65	. . .	. . .	. . .
Stratus	583	2050	120	503	. . .	. . .

[1] See Annals Harvard College Observatory, XXX, 1892, 262.

Kind of Cloud.	Upsala, Sweden. Summer Height.			Storlien, Sweden. Summer Height.		
	Mean.	Max.	Min.	Mean.	Max.	Min.
	Meters.	*Meters.*	*Meters.*	*Meters.*	*Meters.*	*Meters.*
Cirrus	8878	13376	4970	8271	10419	6148
High Cirro-stratus	9254	11391	6840	. . .	. . .	. . .
Low Cirro-stratus	5198	5657	4740	. . .	. . .	. . .
Cirro-cumulus	6465	10235	3880	6337	7358	5233
High Alto-cumulus	5586	8297	4004	4562	4918	4174
Low Alto-cumulus	2771	3820	1498	2744	3844	1182
Strato-cumulus	2331	4324	887	1788	2830	638
"False Cirrus"	3897	5470	2465	. . .	. . .	. . .
Cumulo-nimbus (top)	2848	5970	1400	2504	3515	2998
Cumulo-nimbus (base)	1405	1630	1180	. . .	. . .	. . .
Cumulus (top)	1855	3611	900	2181	2997	1146
Cumulus (base)	1386	2143	743	1401	1901	929
Nimbus	1527	3700	213	1664	5741	617
Stratus	623	994	414	998	. . .	. . .

The prevalence of the different cloud forms at certain altitudes is more clearly brought out by the following table from the same source as the preceding one.

Mean Heights and Velocities of Different Cloud Forms.

Blue Hill, Mass.
(Clayton & Fergusson.)

		1° Cirrus Level.	2° Cirro-cumulus.	3° Alto-cumulus.	4° Cumulus.	5° Stratus.
Summer half year.	Height (meters)	9757	8228	4228	1657	563
	Velocity (met. per sec.)	28.0	24.1	11.2	8.9	7.2
Winter half year.	Height	8012	5039	3484	1571	454
	Velocity	43.9	40.9	20.2	13.7	10.2
Year.	Height	8884	6633	3856	1614	508
	Velocity	35.9	32.5	15.7	11.3	8.7

Berlin, Germany.
(Vettin.)

		1° Cirrus Level.	2° Cirro-cumulus.	3° Alto-cumulus.	4° Cumulus.	5° Stratus.
Summer half year.	Height	—	4520	2395	1310	545
Winter half year.	Height	—	3670	1975	1000	440
Year.	Height	7220	4020	2260	1190	490
	Velocity	17.1	14.0	11.7	10.7	10.2

Although the heights of the several cloud levels vary from summer to winter, the ratios of the successive heights are essentially constant for the year, as appears from the following numbers: the height of the stratus level being taken as unity for each half year.

		Level 1.	2.	3.	4.	5.
Ratios of cloud levels at Blue Hill, Mass.	Summer . . .	1	3.0	7.5	15.0	17.3
	Winter. . . .	1	3.4	7.7	11.1	17.6

The height attained in summer, especially by clouds of middle and upper levels, is generally greater than in winter.

213. Observations of clouds. Weather records should include a statement of the kind and quantity of clouds seen at the usual hours of observation, with their direction and relative velocity of movement. If it is desired to determine simply the relative frequency of clear and cloudy weather, little trouble need be taken to classify the clouds, as their nomenclature is a matter of difficulty because of the frequent occurrence of forms which the observer cannot surely refer to any class; but if the observer wishes to learn something of atmospheric processes for himself, he should give at least as much time to cloud observations as to all the other records together. The various kinds of clouds should be carefully distinguished; the changes from one form to another should be noted, and the relation of the various forms to weather changes should be thoroughly studied out. Descriptive accounts must be often added to the brief terms by which clouds are named; sketches and photographs are of much service in giving definiteness to the record. Instruction in this subject is difficult from the great variety of cloud forms; it can be best gained by reference to photographs or well-executed plates, as verbal descriptions are generally insufficient. It must frequently happen that observers taught only from books will use different names for clouds of the same kind.

The quantity of each kind of cloud should be determined separately; the cloud area is estimated in tenths of the sky occupied by the clouds. In general descriptions of the weather, less than $\frac{3}{10}$ is called clear; from $\frac{3}{10}$ to $\frac{7}{10}$, fair; more than $\frac{7}{10}$, cloudy; overcast is often used to denote an even and complete covering of the sky by clouds of any kind. Cloudless is more emphatic than clear, and refers to a sky in which no clouds are seen.

The direction and velocity of cloud movement are commonly estimated by eye; as "slow from the NW," or "fast from E." It is desirable that a more accurate method should be introduced; for the direction of cloud movement is remarkably steady for hours together and is susceptible of measurement within a few degrees; and the velocity of cloud drifting is certainly a very important element in weather changes. The direction is best determined by noting the path of the reflection of the cloud in a horizontal mirror, at which the observer

looks through an eye-piece that remains fixed during the observation. If the eye-piece is placed so that the reflection of a certain part of the cloud falls at the center of the mirror, and after a few seconds a radial arm is turned so as to bring its edge on the position then taken by the cloud, the edge of the arm will lie parallel to the cloud's motion, on the admissible assumption that the cloud is drifting in a horizontal plane. After setting the arm, it is well to wait again a few seconds to see if the cloud reflection travels along the line thus marked. Without such instrumental aid, the direction of clouds under 30° altitude cannot be safely taken; but with a mirror the directions can be well determined even down to ten degrees from the horizon.

If the eye-piece through which the observer looks at the cloud reflection is always held at a certain height over the mirror, then the relative velocity of the cloud drifting can be measured by counting the number of divisions on the radial arm passed over by the cloud in a given time, as ten seconds. This provides a simple and uniform scale for record, much more closely comparable at different times and places than mere estimate by the unaided eye.

Assuming that each cloud of a class lies at the level determined for others of its kind, the estimates of velocity here described can easily be reduced to actual velocities. Measurements of this kind are strongly recommended to interested observers.

In studying the movement of clouds, it is desirable to discriminate between the generally slow structural movement of one part of the cloud with respect to the rest, and the more rapid bodily drifting of the whole cloud in the air currents. Ordinary records refer only to the latter movement. It is important to note also the manner in which the margin or filaments of a cloud grow or dissolve; and the process by which a cloud changes its form from one class to another. The suggestions of Section 204 should be borne in mind in this connection.

214. Sunshine records. An automatic record of the amount of sunshine — the converse of the amount of cloudiness — is obtained by various instruments. Some employ a sphere of glass by which the sun's rays are focused on a curved sheet of paper at all hours of the day; others secure a photographic print of the track of a fine solar ray that falls through a minute aperture on sensitive paper. In all cases, even a faint cloudiness over the sun may be recognized by a weakening of the record, and heavy clouds covering the sun prevent any record being made. The hours of sunshine for a month, divided by the total number of day-time hours in the month, give an indication of the prevalence of clear or cloudy weather.

CHAPTER X.

CYCLONIC STORMS AND WINDS.

215. Unperiodic winds. In all that has preceded, the reader will find no explanation of the frequent irregular changes of wind and weather. The explanations thus far given account for the diurnal warming and cooling of the air, for the progressive change from winter to summer, for the gradual variation of our prevailing winds from southwest in summer to northwest in winter, for their greater velocity by day and their falling off nearly to a calm at night, for the inflow and outflow by day and night on coast lines, and for the regular diurnal up and down currents in mountain valleys; but all this gives no mention of the shifting of the wind from one direction to another as spells of cloudy and rainy weather pass by, leaving the sky clear behind them. Some additional explanation must be found for the southerly winds of winter that spring up after a time of calm, bringing cloudy skies and rain, often increasing in velocity after sunset, and even causing a rise of temperature during the night; followed perhaps the next morning by winds veering to the west and northwest, when the sky clears and the temperature rapidly falls, even without the customary rise at noon. These quickly-shifting winds belong with the stormy disturbances of the general circulation, to be considered in this chapter. As they spring up at irregular intervals, being brief and light or longer and stronger as may happen, they were referred to a group of stormy winds in the classification already given in Section 138.

216. Cyclones, thunder storms, and tornadoes. Winds of this group are intended to include all those whose causes cannot be clearly referred to some periodic change of temperature, and which are yet certainly dependent directly or indirectly on differences of atmospheric temperature controlled by the sun. We may now subdivide them, taking as the first class those large disturbances so commonly shown on the weather maps as areas of low barometric pressure, 500 or 1000 miles in diameter, accompanied by shifting winds, great areas of cloud and smaller areas of rain or snow; these will be called cyclones[1] or cyclonic storms. The second class includes those smaller disturbances, consisting of cloud masses, ten or a hundred miles in length, advancing broadside or obliquely across the country, giving forth drenching rain, and known as thunder storms from their electric display. A third class includes the violent whirlwinds of excessive destructive force, accompanied by a

[1] See note to Section 266.

pendent funnel cloud from a much greater cloud mass above; these, although commonly called cyclones in this country, will here be referred to as tornadoes. Cyclones, thunder storms, and tornadoes, when fully developed, all possess active or violent winds, and are therefore often referred to under the general word, storms. They all develop clouds so rapidly that rain or snow falls from them, thus causing much more benefit than the occasional harm that is caused by their winds. All these irregular winds possess the common feature of progression as a whole from place to place, being in this respect unlike all other classes of winds.

Beginning with the largest of these disturbances, it should be noted that their more characteristic examples, which will be called cyclones or cyclonic storms, may be associated in a graded series with disturbances of less and less violence until the distinguishing features of their class are hardly perceptible. Their winds may be gentle, their central low pressure faintly developed, their cloud area small and their rainfall practically wanting; and yet under the various names of areas of low pressure, barometric depressions, and barometric minima, all these weak disturbances should be classified with the distinct cyclonic storms in which the same features are more fully developed. In this chapter, however, only the stronger examples will be considered; leaving some mention of the others for the chapter on Weather.

It is advisable to divide cyclones into two subordinate classes, and to consider, first, those which originate in the torrid zone near but not over the equator, and which are therefore commonly called tropical cyclones;[1] second, those which are first seen in the temperate or frigid zones, and which may therefore be called extra-tropical cyclones or cyclonic storms. These two classes are closely alike in many respects, and when tropical cyclones move poleward along a curved path into one temperate zone or the other, as is their habit, they cannot be distinguished from the stronger examples of the extra-tropical cyclones; but the two classes are unlike in certain striking features, as well as in the conditions of their formation, and good reason will appear for referring them to different conditions of origin.

TROPICAL CYCLONES.

217. Tropical cyclones are vast whirlwinds, from one to three hundred miles or more in diameter, whose spiral inflowing currents attain destructive violence near the center, turning systematically to the left in the northern hemisphere and to the right in the southern, around a central area of low pressure. They are accompanied by great sheets and masses of clouds, from which rain falls in torrents, while long cirrus plumes spread out above on all sides. In the center of the whirl, where the pressure is lowest, the wind falls

[1] Properly "inter-tropical cyclones"; but the briefer term is commonly employed.

away, leaving a calm, and here the rain ceases and the clouds may dissolve, showing a clear sky overhead; this central space is therefore called the "eye of the storm." The whole cyclone thus constituted moves slowly along a certain rather well-defined path or track obliquely westward and poleward over the ocean from its sub-equatorial beginning towards the temperate zone, gradually turning to an oblique eastward and poleward motion at latitude 25° or 30°. If land is encountered, the storm weakens, and often dies away. The appearance of a tropical cyclone at sea may be described as follows.

218. Approach and passage of a tropical cyclone. In the torrid zone, the ordinary succession of weather from day to day is remarkably constant. The range of temperature, the faint double oscillation of the barometer, the periodic increase and decrease of cloudiness all show a regularity of recurrence that is unknown in our latitudes. If in such a region the barometer is noticed to rise unusually high, or to stand stationary when its diurnal fall is expected, this may be often on land the first sign of a coming cyclone; but at sea, the faint rise of the barometer is preceded by the arrival of a long rolling swell that swings rapidly out from the storm on all sides, so as to herald its coming even three or four days before its arrival.

The faint rise of the barometer is felt on nearly all sides of the storm area, and it therefore marks what may be called the pericyclonic ring.[1] When its highest pressure is reached, the wind commonly fails. Then fine plumiform cirrus clouds are seen spreading over the sky from the quarter towards the storm center, which may then be one or two hundred miles away in the direction of the doldrums; and about the time of the appearance of these clouds the barometer slowly falls and the calm is succeeded by a gentle breeze. The air becomes sultry, and the sunsets take on lurid colors. When first felt, the breeze generally blows five or six points to the right (in the northern hemisphere) of the direction leading to the storm center. All these signs become more marked as the cyclone draws near; the cirrus clouds thicken and become matted together in cirro-stratus form, veiling the blue of the sky; the refraction of sunlight through the ice crystals of the clouds forms halos around the sun or moon, with the orange or red color on the inner and the blue on the outer side of the circle. Later, the mass of clouds becomes so thick as to obscure the sun, and leave the upper air evenly overcast. The winds have freshened by this time, and blow to the right of a low and distant mass of dark cloud; isolated patches of cloud are seen to form at one side, increase in size, and flow in to join the central nimbus mass. The wind increases to a gale, the waves rise on the sea, the dark clouds approach, thickening as they come, and rain begins to fall from them. The storm center may be then fifty or more miles away, advancing slowly with the whole system of whirling

[1] See American Meteorological Journal, July, 1886.

winds at a rate of eight, ten, or twelve miles an hour. The barometer continually sinking, and at last falls rapidly; with this the roaring wind increases to full hurricane strength, the low scud clouds fly before its blasts, the lightning flashes, the rain descends in drenching torrents, cooling the sultry air. All the elements are in uproar; yet only a day or two before there may have been no sign of the coming storm, except the ominous heaving of the sea.

Before the law of storms was learned, many a ship was borne before such a hurricane, with all sails furled or blown away, helpless in the violence of the winds and waves; and when the vessel was at last about to founder, the wind has suddenly weakened to a calm in the eye of the storm; falling from its greatest violence to an almost perfect repose in fifteen minutes or less. The rain ceases, even the clouds may break away, showing the blue sky by day and the stars by night; but the waves still roll and toss, and in even more dreaded form than in their regular heaving before the hurricane; for in the eye of the storm they swing in from all sides, and pitch and heave tumultuously, forming irregular pits and peaks of water which strain a vessel violently, even to leaking and sinking. A few careful records made in the calm storm center while passing over a land station show a peculiar change in the temperature and humidity of the air. Underneath the surrounding heavy clouds, the air is somewhat cooled and held close to its dew-point by the rainfall; yet the air within the calm center has been found to be comparatively dry with a temperature unduly high; but it is not yet known if these features always prevail. The diameter of the calm space may be ten, twenty, or thirty miles, perhaps a tenth or a fifteenth of the diameter of the whole storm; and its duration in passing a given point may vary from half an hour to two hours: the barometer reading in the center may be even less than 27 inches.

As the hurricane on the further side of the central calm approaches the observer, its moaning can be heard in the distance, rising to a portentous roar as it comes near, and then breaking suddenly with as great fury as the hurricane which died away before, but its direction is now the reverse of that of the winds by which the calm was preceded. All the elements of the cyclone now re-appear; the blasts of the wind beat up the waves to their greatest height, the clouds hang low and heavy over the darkened sea, the rain falls again in torrents; and then as the storm gradually moves away, all these signs of its activity weaken. In the course of a day or two, the barometer rises nearly to its usual height, the wind dies down, the waves fall to a long low swell, the lower clouds recede, the lofty cirrus plumes retreat after them, and the sky is left in its accustomed clearness.

219. Law of storms. Until about 1830, there was little knowledge of the systematic courses followed by the winds in cyclones, and ships at sea were at

the mercy of every storm. Near the beginning of this century, Capper of the British East India Company had announced that the storms of the Bay of Bengal were vast whirlwinds; and about 1826, Brandes in Germany approached an understanding of the general bad-weather storms of that country. A few years later, Dové in Germany (1828), and soon after Redfield in this country (1831), gave full demonstration of the systematic rotation of storm winds, and of the regular progression of the whole disturbance from place to place. These early investigators employed a method of study that has since then been generally introduced in preparing our daily weather maps. They gathered observations from as many stations as possible, and charted on a single map all the facts recorded for a certain definite time, as for noon on a certain date; then again for midnight; for the next noon, and so on; thus producing what are called synoptic maps, and gaining from them a series of views of actual atmospheric processes over a large region. Facts that can be with difficulty perceived from observations at a single station were brought clearly to sight by combining the records of many stations; but while the preparation of modern synoptic maps is simplified by the prompt telegraphic concentration of numerous systematic observations made by trained members of a weather service, as will be more fully described in Chapter XIII, Dové and Redfield had a most laborious task in searching out the scattered records of observers who followed no uniform plan, and who made no regular reports to any central office.

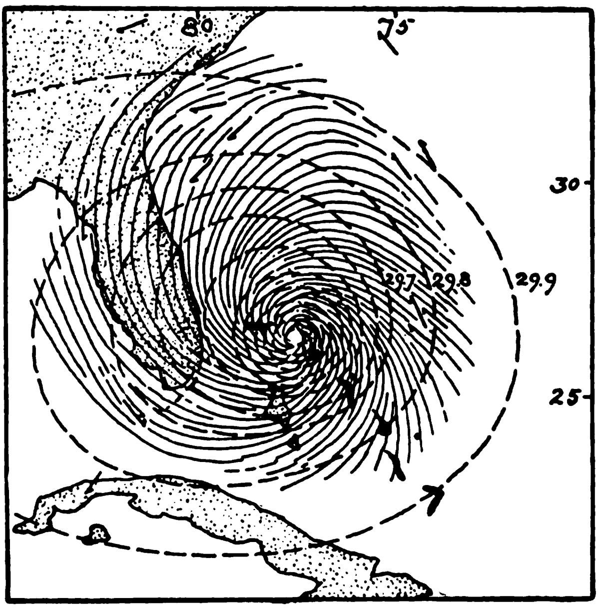

Fig. 53.

Rules based on these discoveries were soon framed, by which the mariner may generally avoid the more dangerous central hurricane winds, and even utilize the more moderate marginal gales to hasten on his way. At first, the surface winds of cyclones were thought to move in circles around the center; but it has since been shown that their incurvature from a circular course in

the outer part of the whirl, and particularly in the rear of the storm area, may amount to as much as two or three points — 23° to 34°, — although close about the center the movement of the wind is much more nearly circular. The isobars and vorticular winds of a violent hurricane on the coast of Florida at Greenwich noon, August 22, 1887, are illustrated in Fig. 53. The steamer "Knickerbocker" passed through the center of this cyclone in the evening of August 23; in the afternoon, the wind blew a hurricane from the east, with

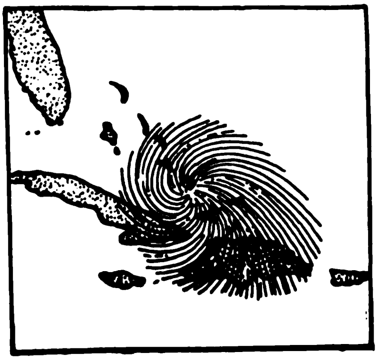

FIG. 54*a*.

FIG. 54*b*.

heavy rain, the sea a mass of foam; at 9 P.M., the wind suddenly lulled; at 10.15, the wind suddenly came out of the west-northwest with fearful violence, terrific rain, and the sea was again lashed to foam; the next morning, the wind moderated with rising barometer, as the vessel steamed southward and the storm moved northward.

The spiral character of the winds is again indicated in Fig. 54*a* and *b*, showing the winds of a Cuban hurricane on September 3 and 5, 1888; the increase in the size of the storm between these two dates is noteworthy. From examples like these, a graphic index, Fig. 55, has been prepared by our Hydrographic Office, in which the winds of any given direction are indicated in their average position with respect to the storm center; all those from the same compass point being on the same dotted line. But while the surface winds turn to spiral courses, the lower clouds follow nearly circular paths around the storm center, and a line at right angles to their movement leads almost directly to the region of greatest danger. Knowing the average size of the cyclonic area and the signs by which its coming is heralded; knowing that

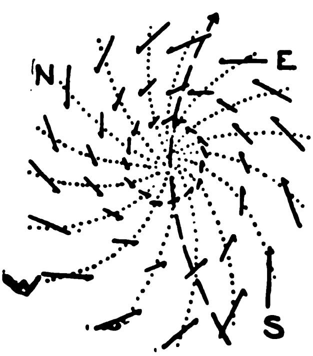

FIG. 55.

the cyclonic winds always whirl to the left in this hemisphere and to the right in the other; and remembering that cyclones move at a moderate velocity westward and poleward while still within the meteorological tropics, and slowly northward as they pass to higher latitudes, it is possible to perceive their coming betimes, and to turn aside from their dangerous centers, thus greatly diminishing the dangers that they bring.

The essential elements of the rules laid down for mariners are: first, avoid running before the wind, particularly on the right of the track (left in the southern hemisphere), for this may lead the vessel along the incurving path towards the storm center; second, lie to on the starboard tack (port tack in the southern hemisphere); that is, trim the sails so as to take the wind on the starboard beam or quarter, for in this way the vessel will necessarily sail away from the storm center into quieter weather. But it should be added that no formal rules will replace the good seamanship that comes from experience, or save a vessel from the many unlooked-for dangers of a storm.

The most dangerous portion of a cyclone lies somewhat forward and to the right of its center in this hemisphere, to the left in the other; for here the winds are usually most violent, here the advancing center of the storm may overtake a vessel that is attempting to run forward and cross its track, and here the curvature of the path of the storm constantly brings the violent center closer to the ship. Vessels taken unaware in this dangerous quarter of a cyclone may be unable to escape its violence.

A modern addition to the older rules for diminishing the danger of storms at sea is to spread oil on the waves, whereby their height is lessened and they break less frequently over the vessel. Even a small quantity of oil allowed to drip from a bag hung over the vessel to windward has been found by repeated experiment to be of great service.

220. Evidence of convectional action in tropical cyclones. In attempting to explain these violent commotions of the atmosphere, we must resolve the motion of the wind, *AB*, Fig. 56, into two components, one of which, *AD*, is directed radially inward toward the center, *C*, while the other, *AE*, runs around the center with increasing rapidity as the center is approached. In the outer part of the cyclone, the radial component is almost equal to the circular; near the center, the circular component is much the greater of the two. Indeed, if the circular component of the wind's motion were absent, and the air moved simply as a radial inflow instead of in a spiral whirl, its velocity would be so moderate that it would not reach the violence of a storm wind. Different causes control the two components of the motion of the wind; we shall first look for the cause of the radial inflow.

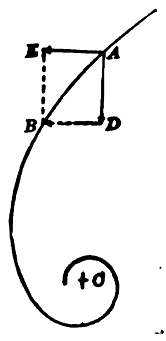

FIG. 56.

Inasmuch as the central barometric pressure remains low, or, as the storm grows stronger, even becomes lower than before, in spite of the gradual spiral inflow of the winds, some escape for the air from the central region must be inferred. An upward escape is indicated by the occurrence of the great cloud mass above and by the drenching rains that fall from it; for no cause of an extensive condensation of vapor is so effective as the mechanical cooling of ascending currents. The upward escape is in turn confirmed by the outward spreading of the cirrus plumes in all directions aloft. There can be little doubt that the air, which slowly approaches the center below as it whirls around in rapid circuits, finds an escape by ascending gradually in the area around the central calm, whirling as it rises and flowing spirally outward far above sea level. The lower warm damp air, cooling a little by expansion as it advances towards the central lower pressure, and by the fall of rain from the clouds above, forms the lowest scud clouds flying before the wind. Cooling much more in its spiral ascent around the center, it forms the heavy cloud mass from which fall the drenching rains that always accompany tropical cyclones. The last of the condensed vapor, excluded at the highest levels at temperatures below freezing, takes the form of ice crystals and makes the long outflowing cirrus plumes. Beyond the extremity of these clouds, it is probable that a gradual descending motion of the outflowing air takes place in the pericyclonic ring of higher pressure and clear dry air, by which the cyclone is surrounded.

The indications of inflow, ascent, outflow and descent in systematic order thus suggest that the inward component of the surface winds is one member of a large convectional circulation, such as was explained in Chapter VI; while the whirling component is produced in some other way. If the convectional origin of the inflowing surface winds is accepted, their movement must be ascribed to the inferred settling down and creeping in of heavy marginal air; for convection always depends primarily on a descending motion under the pull of gravity. But before the action of gravity can produce a descending movement in the atmosphere, there must be an excess of heat localized in the area which afterwards becomes the storm center with its ascending currents; and this seems to be the fact. In all cases where tropical cyclones have been traced backwards towards their source, they seem to have come from near the warm equatorial regions, where the accumulation of over-warm and moist air is a characteristic feature. Yet if such is their origin, they should be expected at all seasons, for the middle torrid zone has a continual summer and an ever-ready instability. The following sections will show, however, that this is not the case, and will also suggest a sufficient reason for the limitation of the occurrence of tropical cyclones to particular seasons.

221. Seasons and regions of tropical cyclones. There are five regions in which tropical cyclones appear with characteristic regularity. The first to be mentioned is in the North Atlantic, where they are thought to begin at a distance of six or eight degrees north of the equator, and somewhat towards

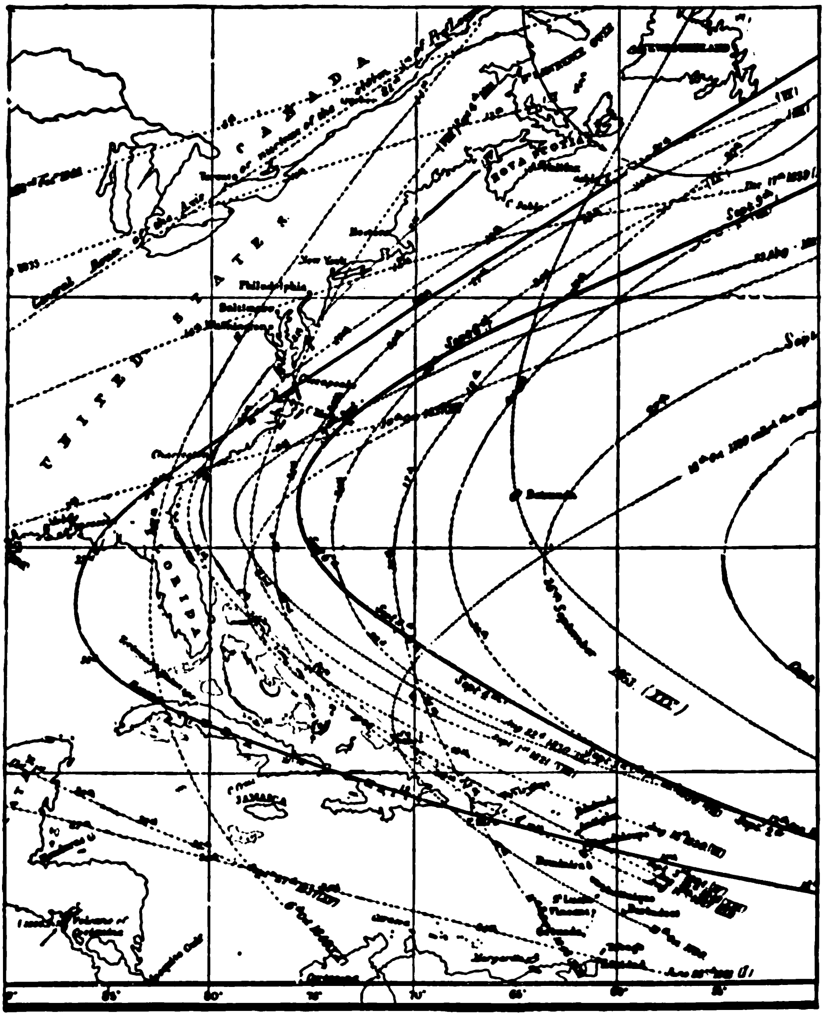

Fig. 57.

the warmer western side of the ocean. The cyclones are small when first observed, with almost radial winds, but they increase in size and violence and rotary motion as they move westward, then northwestward over the Lesser Antilles, still recurving to the right until they advance directly north for a

short distance opposite Florida, and then turn off to the northeast along or near our coast, frequently continuing as far as northwestern Europe before they fade away. It was in the study of these cyclones from 1830 to 1850 that Redfield contributed so much to the law of storms. Fig. 57 gives the tracks of many Atlantic hurricanes traced by Redfield and Reid, reduced from one of Redfield's latest essays, published in 1854; but it is probable that the curves are here too regularly drawn. This map forms the basis for the indications of storm tracks on the Pilot Charts of the North Atlantic, issued monthly by our Hydrographic Office at Washington.

A special account has been issued by the Hydrographic Office of the St. Thomas-Hatteras hurricane of September 3 to 12, 1889. It was felt on the Windward Islands on September 2. On the 3rd, it had a diameter of about 500 miles with a calm center about 16 miles across, and thus passed close to the north of St. Thomas. On the 4th, it was north of Puerto Rico, where northerly, westerly, and southerly gales were successively felt; the clouds of the cyclone were seen on this day from Turk's Island, causing alarm until they drifted away to the north. On the 5th, the storm center lay about 300 miles north-northwest of Santo Domingo; its enormous waves had overspread the greater part of the western Atlantic in a heavy swell, felt even at Nantucket. On the 6th, it was midway between Cuba and the Bermudas, still having violent winds and heavy clouds and rain. By the 8th, it lay about 300 miles east-southeast of Cape Hatteras, and a northeast gale blew along our coast from Maine to Carolina; on this and the subsequent days, great damage was done by the surf on the New Jersey coast. The storm moved slowly northward, and after the 10th, when the center was off Norfolk, its winds weakened, and on the 12th its fury was exhausted. Hundreds of vessels that had been storm-bound in our harbors set sail in the fair weather that followed. Fig. 58 represents the great cyclone of November, 1888, on the 25th of that month, on its way along our coast. Fig. 106 shows the isobars and gradients of the western half of the same storm on November 28.

The season at which tropical cyclones are observed in the North Atlantic is limited to the late summer and early autumn; the months in which they are commonest being August, September, and October, while they are practically unknown from December to June. Their annual number seldom exceeds six or eight; and only a few of these may reach the greatest violence.

In seeking a cause for their coming to the West Indies in the months about the autumnal equinox, it is noticed that in these months the equatorial calms or doldrums of the Atlantic migrate farthest north of the equator, and that in tracing the cyclones backward along their track, it is in the calm region of warm, moist air between the steady trades that the apparently convectional overturning of the West Indian cyclones has its beginning. Such a region of quiet air under strong sunshine is the natural seat of the most

pronounced convectional action. The air being quiet becomes warm and well moistened by evaporation; the warm and moist air becomes unstable and takes on a gradual convectional overturning, and from this beginning the development of the cyclone is thought to proceed.

In the same way, the western and warmer equatorial Pacific southeast of Asia furnishes to the Philippine and Japanese islands and the neighboring

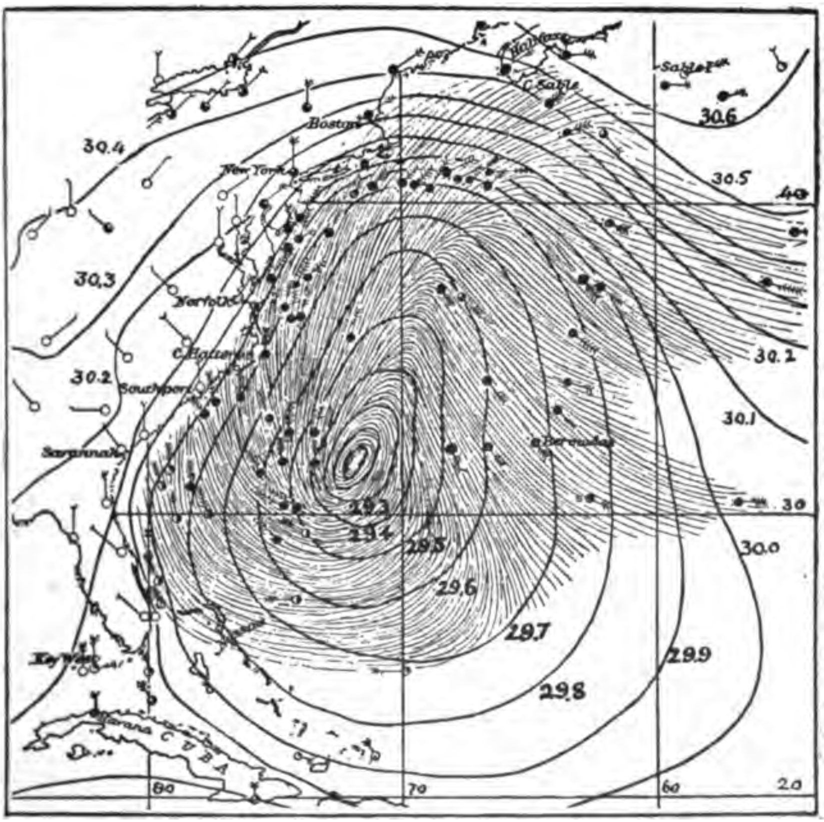

Fig. 58.

coast of China a succession of cyclones, there called by the Chinese name of typhoons, in the months of August, September, and October nearly every year, when the Pacific doldrums are farthest north. It was concerning one of these, crossing the Philippine islands in 1882 on its usual northwestward course for that latitude, and making automatic record of its weather in the meteorological observatory at Manilla, that the statements were made above regarding the high temperature and low humidity in the calm eye of a cyclone. Certain

cyclones that have been traced for a relatively short distance along a northwestward course in the north torrid zone of the Pacific ocean west of Central America are also supposed to have originated in the equatorial calm belt. They are not known to cross the trade wind belt.

Again, in the southern Indian Ocean, the islands of Mauritius and Reunion are annually visited by a series of cyclones, here coming from the northeast, recurving in latitude 25° or 30° south, and then passing off southeastward to the south temperate regions. It was from the study of these that Meldrum, the meteorologist of Mauritius, was among the first to announce the true incurvature of cyclonic winds from their supposed circular course. In the northern hemisphere, the cyclone season occurred when the doldrums stood farthest north of the equator; in this southern ocean, they spring up when the doldrums move to their farthest southern latitude; that is, in February and March; thus giving additional confirmation of their convectional origin in the belt of equatorial calms.

Cyclones are little known in the South Pacific ocean, but are occasionally met with in the region east of Australia, where the ocean is warm and where the trade winds are weak. Their season of occurrence is in the later summer or early autumn of the southern hemisphere. It was in this region that a hurricane in March, 1889, wrecked or injured many naval vessels belonging to different nations at Apia, Samoa; and that a hurricane in March, 1886, was severely felt on the Fiji islands.

Cyclones occur with dreaded violence in the northern gulfs of the Indian Ocean, particularly in the Bay of Bengal. It was in the study of these storms that Piddington first proposed the name cyclone, fifty years ago; and here in later years the Meteorological Service of India has gathered the fullest information now in hand concerning the early stages of cyclonic action. The violent cyclones of the Bay of Bengal are unlike those of other parts of the world in occurring in two seasons instead of in only one. They occur in the southern and central parts of the Bay first in April, May, and June, and again in October and November. In the intervening summer months, less violent cyclonic storms are formed in the north of the Bay and on the land. Their general progression is to the northwest, but they turn more to the north or even to the northeast in the earlier and later months of the year. Good reason for the double season of occurrence of the more violent cyclones is found in the prevalence of calms over the Bay of Bengal in these two seasons of the year: first, when the sun advances northward, and again when it returns southward; while the cyclonic storms of midsummer are formed farther north, but often on land, corresponding to the northernmost position of the heat equator. This gives still further ground for ascribing tropical cyclones to the calm areas of the migrating equatorial belt.

222. Tables of cyclone frequency. The tropical cyclones in the four chief regions of their occurrence have been tabulated by various meteorologists to illustrate their distribution through the year, as appears in the following lists. The numbers are not strictly comparable, for no precise standard of violence is yet adopted to determine whether a storm shall be counted or not. In most of the examples, the storms have not been traced backward along the track to their source.

LOCALITY. — AUTHOR.	PERIOD.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	YEAR.
West Indies.														
Poey	1493 to 1855 .	5	7	11	6	5	10	42	96	80	69	17	7	355
China Seas.														
Schück	85 years . .	5	1	5	5	11	10	22	40	58	35	16	6	214
Bay of Bengal.														
Blanford . . .	To 1876 . .	2	0	2	9	21	10	3	4	6	31	18	9	115
Eliot	1877 to 1891 .	–	–	–	–	10	17	27	24	28	18	21	7	152
Arabian Sea.														
Chambers . . .	To 1881 . .	4	3	2	9	13	20	2	2	3	4	10	2	74
Eliot	To 1888 . .	2	0	0	8	9	5	0	0	1	3	8	1	37
S. Indian Ocean.														
Meldrum . . .	1848 to 1891	63	76	49	33	11	1	1	0	0	2	14	29	279
		19	6	19	24	11	2	1	0	0	3	13	10	108

The list of cyclones for the West Indies includes all the violent storms of that region mentioned by older and newer authors down to 1855. It is highly probable that a number of storms noted in the winter months do not really belong to the class of tropical cyclones. A tabulation has been made for this region in recent years by Finley.

The figures for the Bay of Bengal for 1877 to 1891 include many cyclones of moderate size and intensity that formed in the north of the Bay or over the adjacent land, where they were detected on the daily weather maps of India. These do not exhibit a double period such as appears in the earlier list, which includes only the larger and more violent cyclones, whose origin was in the central or southern part of the Bay. The late winter storms of the Arabian sea included in the list by Chambers probably originated, at least in part, in temperate latitudes north of the sea, and should not be included with the rest as of tropical origin.

The data for the South Indian ocean have been especially furnished by Mr. Meldrum of Mauritius. The first series of figures includes those cyclones whose progression has been well determined; the second series gives the storms which have not been shown to change their position, these being generally of brief duration.

223. Early stages of cyclonic action. In most of the oceans the early stages of cyclones have not been fully observed; but in the Bay of Bengal, where cyclones are relatively numerous, the logs of many vessels passing to and fro have been carefully examined for the fortnight before the occurrence of storms, and thus the conditions of their beginning have been well determined. The most notable antecedent condition observed before the appearance of a cyclone in the Bay and presumably occurring also in other regions of cyclone growth, is the uniformity of pressure and the quietness of the air over the sea. There may be light local breezes, but there is no persistent movement of the atmosphere such as prevails during the occurrence of the summer or winter monsoon, when the winds sweep steadily across the whole breadth of the waters. The quiet air becomes over-warm and moist, and clouds hide the sky in the calm region; with this, the pressure decreases slightly, and gentle marginal breezes are established towards and around the central district. Rain sets in under the central clouds, the barometer falls to lower readings, the winds blow stronger and with more definite courses, all conspiring to form a vorticular whirl about the center of lowest pressure. It is probable that the irregularity sometimes noticed in the winds at the inception of the cyclone results from the imperfect development of several low-pressure centers; but at a little later stage, one of these alone survives and becomes the eye of the cyclone, and the isobars assume an almost circular form around it. As the pressure falls towards its lowest value and the winds attain their greatest strength, the center of the cyclone advances from its first vague position along the usual northwest course. It progresses with growing violence until it reaches the land; then its on-shore winds sweep the waters of the Bay over the low delta plain of eastern Bengal, where the inhabitants have thus been drowned by the tens and almost by the hundreds of thousands. Further inland, the storm generally weakens; and on approaching the mountains to the north or the elevated plateau country in the south of the peninsula, it fades away. The early stages of storm growth are here so well observed and indicate so clearly a convectional beginning for the cyclone that this theory of their origin here and in other parts of the torrid zone is now generally accepted.

While the foregoing paragraphs give good reason for associating the formation of cyclones with convectional action in the equatorial calms or in the weaker part of the trade winds, no full explanation has yet been given for their limitation to special seasons of occurrence, when the calms are furthest from the equator. The conditions for convectional overturning are always present in the equatorial calm belt in a greater or less degree; the air there loiters about, moving gently in light baffling breezes, reaching as high a temperature as is found anywhere over the ocean, always well moistened by evaporation from the sea, and ready to ascend whenever cooler, heavier air

flows in beneath it. Some additional reason besides instability must therefore be found to limit the development of convectional cyclones to that season when the calms migrate on the sea surface farthest from the equator, either into the northern or southern hemisphere. The reason sought for may be found when it is remembered that a direct convectional indraft of the surface winds cannot alone produce any whirling motion, and that the terrific blasts of the wind near the center of tropical cyclones are always in an almost circular path. Cyclones must therefore be essentially vorticular storms, and although begun by convectional action, some supplementary cause must set them in rotation.

224. Effect of the earth's rotation. Tropical cyclones being essentially vorticular storms, a sufficient explanation of their occurrence only when their place of origin is removed from the equator must have already come to mind from the account of the deflecting action of the earth's rotation given in Chapter VI. Instability and convection occur in the doldrums at all seasons, as shown by the numerous thunder storms and heavy rains of the calm belt, but the establishment of a convectional whirl with a definite direction of rotation can take place only when the doldrums are far enough from the equator to give the deflecting force an effective value and allow it to require the winds to depart systematically from directly radial lines of inflow. The departure of all the inflowing winds being to the right in this hemisphere, or to the left in the other, they must all conspire to produce a left-handed whirl if they begin north of the equator, or a right-handed whirl if they are developed south of the equator. The origin of the circumferential component of the wind's motion, *AE*, Fig. 56, is thus accounted for.

Recalling the explanations of Section 133, it will be understood that any mass of calm air in the doldrums or elsewhere may be compared to a paper disc attached to an artificial globe; or to a vessel of water on a turning table; and hence that while its parts are quiet with respect to the surface of the sea on which they lie, they nevertheless possess a movement of rotation with respect to their center in consequence of their residence on a rotating planet. The air at the equator corresponds to a stand-still of the turning table. The air in the calms north of the equator corresponds to the water in the vessel when the table is turning slowly from right to left; south of the equator, when the table is turning from left to right. The air in temperate latitudes corresponds to a faster rotation of the table. If these relations are appreciated, there can be no difficulty in explaining the limitation of tropical cyclones to certain seasons, when the calms in which they begin have migrated far enough north or south of the equator.

Recalling next the experiments with the eddies of water in the rotating vessel (Sect. 135), the violence of the cyclonic whirl may be appreciated; it being understood that while the water eddy is discharged downward, the

atmospheric eddy is discharged upward. It has been seen that if the water is discharged after it has been given a gentle rotation, the outflowing currents soon develop a violent central vortex, where the velocity of the threads of water is much greater than the velocity of rotation at the margin of the vessel, and very much greater than was seen at any part of the discharge when there was no rotation. The centrifugal force developed by the rapid whirling of the water on a small radius produces a distinct depression of the water surface at the center; the whirl may become so violent as to form an empty core as a result of the excess of the horizontal centrifugal force in the whirl over the downward action of gravity. Finally, the discharge of rotating water requires more time than the discharge of quiet water; from having taken a quarter of a minute when quiet, it may occupy forty or fifty seconds when rotating.

These simple experiments illustrate motions that are analogous to the flowing of the surface winds in the convectional overturnings of the equatorial calms, with the respective parts inverted. The top of the water corresponds to the bottom of the atmosphere. The downward discharge of the water from the vessel corresponds to the convectional ascent of the air in the doldrums. The direct discharge of the quiet water illustrates the simple convectional overturning of the air, when the calms are near the equator and no cyclonic whirls are produced. The whirling escape of the rotating water represents the whirling inflow of the winds when the calms stand far enough from the equator for their breezes to be governed by the deflective action of the earth's rotation. Moving gently inward at first, the whirling velocity continually increases as the center is approached, and the wind attains a full hurricane violence close around the area of the central calm, where it follows an almost circular path.

The origin of tropical cyclones thus appears to be well worked out. Being of convectional nature, they are not formed in the steadily-moving trades, but only when the trades weaken in the loitering doldrums, where the lower air becomes excessively warm and moist; being essentially whirling storms, they cannot develop when the doldrums are close to the equator, where the inflowing currents are not required to unite in forming a systematic vortex; but only when convectional action begins at some distance north or south of the equator. The definite and rational association of these various conditions of storm growth gives warrant for much confidence in the convectional theory of the formation of tropical cyclones.

225. Absence of tropical cyclones from the South Atlantic. All the torrid oceans are visited by tropical cyclones, except the South Atlantic. They are not common in the broad South Pacific; but they have been observed there and with terrible violence, as at Samoa in 1889. But the south torrid zone of the Atlantic has no record of true cyclones. The reason for this peculiar exception is apparent when the attitude of the doldrums in January and February is

examined in Fig. 59. The shaded areas in this figure represent the location of the equatorial rain belt for the northern and southern positions of the heat equator. The trades die out as they enter this belt, their limits being indicated by broken or dotted lines, and the calms of the doldrums are found between these limits. The calms migrate about ten degrees northward in our late summer, and there give forth cyclones that travel off toward the West Indies; but they never migrate so far south of the equator, being held back by the great current of relatively cool water that advances from the Antarctic ocean along the west coast of Africa, as has already been dwelt upon in describing the distribution of temperature (Sect. 82). The South Atlantic is therefore the only ocean into which the doldrums do not migrate. It is also the only ocean not visited by tropical cyclones.

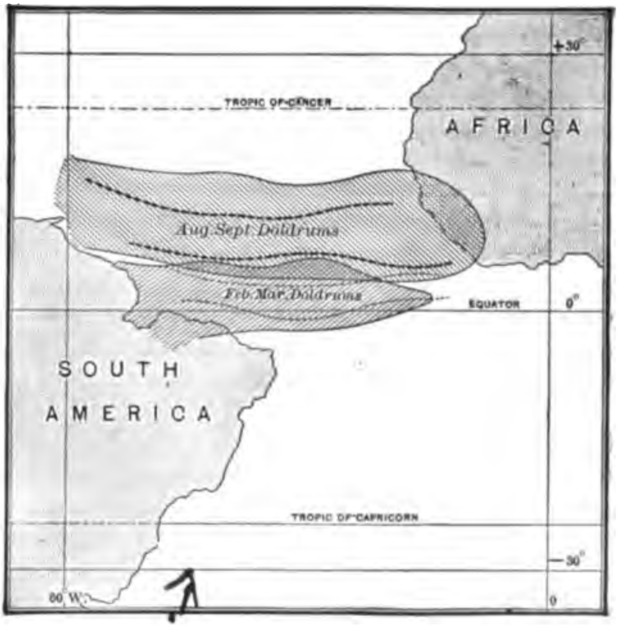

FIG. 59.

226. Latent heat from rainfall. When the cloud mass of a cyclone is formed and its rainfall has begun, then not only the sensible heat of the warm air but the latent heat of the condensed vapor promotes the convectional action of the storm. The consideration of this double process has already been given in the chapter on clouds: here one of its most important applications is encountered. As the warm air ascends in its spiral whirl around the central region of a cyclone, it must expand; and at first it draws only on its sensible heat for the energy needed to push away the surrounding air. Its temperature then falls at the rapid rate of 1°.6 for every 300 feet of ascent; but as the whole mass is very damp, a moderate ascent is sufficient to reduce the temperature sufficiently to overtake the falling dew-point and cause condensation. At this moderate altitude the clouds begin to form. Expansion continues during ascent to greater heights, but the energy for expansion is

then drawn from two sources; a part comes from the sensible heat of the ascending air, and the remainder from the latent heat of the vapor that is condensed in the ascent.

Here, as in all cases where vapor is condensed, it may be looked on as giving up the store of energy that was acquired from absorbed insolation when the vapor was formed, perhaps many days before and hundreds of miles away. In the case of tropical cyclones, the store is abundant and its aid is most effective. The cyclonic inflow comes at first from the doldrums, and afterwards from the trades when the storm area increases and when its progression carries it to higher latitudes. The air thus supplied has a high temperature and a high relative humidity. Enormous masses of heavy clouds are formed, and rain falls from them in drenching torrents. The diagram in Section 197 has shown that it is precisely under these conditions that the liberation of latent heat causes the greatest retardation of cooling in an ascending current, and that a convectional ascent may reach its greatest height. The greater the altitude of the cyclonic mass in which the temperature is higher than that of the surrounding air, and the greater the excess of the central temperature, the stronger the gradients and the more violent the winds. As both the excess of central temperature and the altitude of ascent are greatly promoted at high temperatures by the condensation of vapor and the liberation of its latent heat, it is manifest that the presence of vapor is a very important element in the development of tropical cyclones.

In order to realize the enormous amount of energy needed to develop a tropical cyclone, we may quote a comparison that has been drawn between such a storm and a large ocean steamer. The air in a cyclone 100 miles in diameter and a mile high weighs as much as half a million 6000-ton ships; and yet this enormous mass is set in rapid motion, averaging over 40 miles an hour, in the course of a few days, and its motion may be continued for a week or more. Again, the Cuban hurricane of October 5–7, 1844, is calculated on very moderate estimates to have worked during the three days of its progress along our southern coast with an energy of at least 473 million horse-power. The continued maintenance of so enormously powerful a disturbance calls for the rapid supply of a vast amount of energy; just as the active steaming of a large engine calls for a plentiful supply of coal under its boilers. In the case of a fully-developed tropical cyclone, it is believed that the energy is chiefly supplied from the latent heat of the heavy rainfall; and reasonable estimates of the amount of condensation within the storm disc show that this source of energy is ample in amount.

227. Occurrence of tropical cyclones chiefly over the oceans. The torrid lands of Africa and South America are not, as far as observation goes, visited by tropical cyclones. The insular and peninsular lands of southern Asia are

reached by cyclones that come from the adjacent seas, but after leaving the ocean, their violence is greatly reduced, and if they encounter high ground they are broken up. It is argued from this that the development of cyclones in the calms of the doldrums is limited to those parts of the belt which are most highly charged with vapor, that is, to the parts over the oceans; and further that the maintenance of the storms is difficult on the land where the water vapor is in smaller amount, and where the vorticular circulation of the lower winds is more or less interfered with by mountains or plateaus.

228. Comparison of tropical cyclones and desert whirlwinds. An instructive comparison may now be drawn between the great tropical cyclones and the small dusty whirlwinds of desert plains. Desert whirlwinds are slender columns of immediate and local formation and of brief action. Tropical cyclones are broad discs of gradual and widespread formation and of long endurance; their winds and the vapor which is condensed in their clouds may be drawn in from districts several hundred miles away from the violent whirl that is generated around their center; they may last for a week or two, blowing night and day and travelling thousands of miles away from their starting point. The desert whirls spring up about ten or eleven o'clock in the morning, after the steepening rays of the sun have warmed the barren ground and the lower air has been warmed from the ground by conduction and radiation. Some little inequality of the surface or a slight movement received from the winds aloft causes an upsetting of this unstable arrangement of the lower air, and an inflow is thus begun towards the place of ascent; but as the various inflowing currents move for too short a distance to be systematically influenced by the earth's rotation, and as their irregular flow does not allow them to meet precisely at a center, they turn a little to one side or the other according as the stronger inflow decides, and a little whirl is then developed, rotating indifferently one way or the other. As its violence increases, dust and sand are gathered up by the wind and the lofty, slender column becomes visible. It may be followed to a height of several hundred or even a thousand feet, where it spreads out laterally, the coarser sand soon settling down, while the finer dust is borne many miles away. The supply of warm surface air is soon exhausted and the whirl quickly disappears; but within half an hour another layer of surface air may be superheated and a second whirlwind arise from it. So brief a process as this is only a slight exaggeration of the invisible convectional movements on which the diurnal increase of the general winds over the land depends.

The formation of tropical cyclones is much more deliberate. Day after day the air lying quiet over the sea becomes warmer and warmer, the vapor gradually rises by diffusion and by local convection to higher and higher levels, nearly saturating a large volume of warm air. When at last some

overflow of the expanded air is developed aloft, the more general ascent and overclouding begins and a creeping in of the surrounding air is established; but all these changes take place very gradually. The observer only notes a slow increase of cloudiness and a strengthening of the wind as the days pass. While the desert whirl quickly spends itself, the tropical cyclone draws upon so large a volume of air that it lasts many days; and the prompt convection that would drain away the supply, if the inflow were directly radial, is greatly delayed by the systematic deflection of the winds to the right or left of their radial path and the consequent development of a gigantic whirl around the central area of low pressure. While the desert whirl can continue only during the supply of warm air in the hot hours of the day, the cyclone may persist with undiminished energy over night, not only because the temperature of its broad and thick layer of cloudy air is about the same day and night, but also because the greater supply of energy from the latent heat of its condensing vapor is furnished at night as well as in the day-time.

While thus contrasted in many particulars, these two classes of whirls are alike in one essential feature. The winds of both are driven by the gravitative or downward pressure of a surrounding mass of heavier air, which settles down as the whirling lighter air ascends; and the essential cause of the difference in weight of the central and surrounding masses is found in the higher temperature of the former, dependent in some way on a greater absorption of insolation. Here, as in all classes of winds, except those expressly excluded in Sections 139 and 140, the movement of the atmosphere depends on the interaction of solar energy and terrestrial gravitation.

229. The eye of the storm. Mention has already been made of the calm area within the whirling hurricane winds, where the pressure is lowest and where the rain ceases and the clouds sometimes break away, revealing the clear sky overhead. All these features of the center of a strong cyclone seem to be easily explained as consequences of its convectional whirling. As the winds are pushed in towards the central area of low pressure, their velocity of rotation around the center becomes excessive and the centrifugal force[1] increases at a very rapid rate, as explained in Section 135. The low pressure at first caused by high temperature is thus greatly intensified and the gradients become very steep near the center. The central pressure sometimes falls even three inches below the normal value of the region; and it is plain that as the winds approach so rarefied a region, they must expand and cool and become cloudy; it is presumably in great part for this reason that the flying clouds around the eye of a cyclone hang so low over the sea.

[1] Distinction should be made between the true centrifugal force, arising from the whirling of the wind around the storm center, and the deflecting force arising from the movement of the wind on a rotating earth; but as the former is much the greater of the two in tropical cyclones, it alone is named in the text.

The air that is held away from the storm center by the excessive centrifugal force there developed aids the convectional overflow aloft in forming the ring of slightly higher pressure around the storm, already referred to as the pericyclonic ring in Section 218: here the clearness of the sky and the freshness of the air confirm the theoretical suggestion that there is a gentle descent from aloft; and this is further demonstrated by the relative calmness of the air in the ring of high pressure, and by the slow outward spiral movement of the air outside of the ring, perceptible in the charts of the best-studied tropical cyclones.

The inflowing winds take an almost circular course near the center. They are sometimes intensified by brief and violent gusts of terrific strength; and sometimes deflected from their average course by the passage of subordinate eddies; but on the whole, the observations reported from vessels that happen to be near the vortex of a cyclone agree remarkably well in indicating a systematic whirl of the winds around a single center. When blowing in this way, nearly all the strong centripetal force on the steep gradients is used in overcoming the strong centrifugal force of the violent wind on its small radius, and there remains a forward component of moderate value, sufficient only to overcome the small resistances encountered in wave-making and internal friction. At moderate altitudes, where the resistances are smaller than at sea-level, the wind takes a more nearly circular course; hence the movement of the lower clouds is prevailingly a point or two to the right (in the northern hemisphere) of the surface wind. On land the resistances are still greater, and there the storm winds are more nearly radial than at sea.

The approach of the winds to the center of a cyclone at sea is delayed by their having to follow a spiral course; at the level of the low-hanging clouds the winds are essentially circular, and cease further inflow at a distance of ten or fifteen miles from the center of their whirl. In the meantime the convectional ascent of the air around the central region is continued; and before any part of the indraft can enter very close to the center of the whirl, it is carried upward to high levels, and then turned spirally outward at the altitude of the upper clouds. A certain space about the center, commonly measuring ten or twenty miles in diameter and three to five miles high, is therefore inaccessible to violent winds; its air is comparatively calm, although surrounded by winds of hurricane strength. The empty eddy in the center of a vortex of water and the clear core often observed in a dusty whirlwind are analogous to the calm central area of a cyclone: the centrifugal force of the surrounding currents is so great that they cannot approach closer to the center before they are carried away; upward in the air, downward in the water.

The clearness of the eye of the storm has been thus explained: The hurricane winds that surround the central calm area must impart some of their motion to the enclosed air by friction; the air thus given a rotary movement

must thereby acquire a certain centrifugal force, and thus increase its pressure against and its movement with the surrounding hurricane. Some air will thus be withdrawn from the calm center, and carried up in the surrounding convectional whirl. To supply the air thus withdrawn from the margin of the calm center, the quieter air within must expand laterally, and thus allow some air from aloft to settle down to the sea, as indicated in Fig. 60; further-

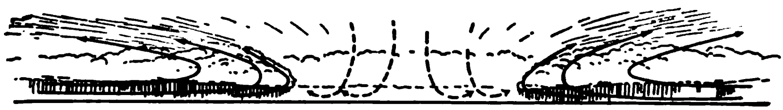

FIG. 60.

more, the depression of the isobaric surfaces close around the center is so great that an inflow may be developed on them in the higher atmosphere, above the level of the whirling storm. If a slowly descending current can be developed in this way, it is natural enough that it should be clear, because it will be warmed by compression, and any clouds that may wander into it will soon be dissolved. For the same reason, if the air descends actively even to sea-level, it should be hot and dry, and in strong contrast to the damp and somewhat cooled air of the surrounding stormy winds. This is not always the case, although it was so to a marked degree in the hurricane of 1882 at Manilla on the Philippine islands, during the passage of the central calm.

It should be noted that while the calmness of the central area may be fully explained on mechanical principles that certainly have application in a whirling storm, the descent of the air in the central space is merely a suggestion, plausible in certain respects and capable of explaining the phenomena for whose explanation it is proposed; but not yet fully verified by observation. Observations of the eye of cyclones while they are passing over islands may in the future serve to decide this question.[1]

230. Comparison of tropical cyclones and the circumpolar whirl of the planetary circulation. The comparison that may be here drawn aids greatly in the understanding of both systems of winds.

1°. In the circumpolar whirl the center is cold, and the air there descends; the high polar pressures expected from low polar temperatures are reduced to low pressures by the excessive centrifugal force of the whirling winds; and the expected gradients towards the equator in the lower air are reversed to poleward gradients, except in the trade-wind belts. In tropical cyclones the central region is warm, and the air there ascends; the low pressure due to high temperature is reduced to even lower pressure by the centrifugal force of

[1] A full account of the features of the "eye of the storm" is given by S. M. Ballou in the American Meteorological Journal, June, July, 1892.

the revolving hurricane; and it must be surmised that the outward gradients which a simple convectional circulation would demand in the upper air, are thus reversed to inward gradients.

2°. The central low pressures thus determined in both systems of whirling winds are accompanied by a partial re-arrangement of the surrounding pressures, producing a ring of high pressures where the air slowly descends, and on whose circumference of no gradients there is a belt of calms and fair weather separating the interior inflowing spiral winds from a set of exterior outflowing spiral winds. The high-pressure ring is seen in the tropical belt of high pressures of the planetary winds, outside of which the trade winds blow in an outward spiral; and in the pericyclonic ring of high pressure in tropical cyclones, beyond which are faint external outflowing winds.

3°. Land masses with plateaus and mountain ranges interfere with the best development of these systems of winds and pressures. When tropical cyclones run ashore, they weaken and often disappear. Similarly, the northern half of the planetary circulation, flowing over the broadest continents, the highest plateaus and the most numerous mountains, is imperfectly developed in comparison with the southern half, where the sea surface is so little interrupted.

4°. The return of the planetary winds from the polar regions is effected against the (apparent) gradient by virtue of the excessive centrifugal force previously developed while the winds were obliquely approaching the pole on the steep gradients aloft. The observed outflow of the cirrus clouds in the upper part of a cyclone may similarly have to be performed against the inward gradient of the upper isobaric surfaces by means of the great centrifugal force that the hurricane has previously acquired as it approached the storm center on the even steeper isobaric surfaces of lower levels.

5°. The observed calmness of the air in the eye of a tropical cyclone lends support to the inferred calmness of the air in the polar regions of the planetary circulation, mentioned in Sect. 142.

6°. The surface members of the planetary circulation, being reduced in velocity by friction with the earth, are unable to return to the equator against the poleward gradients, and hence sidle obliquely towards the poles. The surface members of the cyclonic whirl, being retarded by friction and wave-making to a lower velocity than that gained by the winds at the level of the clouds, are constrained to take a somewhat greater inclination towards the central region, where the pressures are so greatly reduced by the violent whirling of the whole mass.

It is manifest that comparisons of this sort are of a different order from those which are concerned with statistical values, based directly on observation, such as are commonly employed in climatic studies. Each style of comparison has a value of its own; both must be employed if the student would reach an

appreciation of the science as well as a knowledge of the facts of meteorology. Statistical comparisons have been longer employed, and for a time they formed the chief subjects of meteorological study. Comparisons of similar phenomena are less usual, but not less important. The one here introduced was first made by Ferrel, to whom the modern understanding of the theory of the winds is so largely due. It should be carefully studied, for it is as important in meteorology to perceive the homology that exists between the larger and smaller atmospheric whirls as it is in astronomy to understand that the movement of planets around the sun is controlled by a system of forces corresponding to that which directs the movement of moons around planets.

231. The convectional theory of tropical cyclones. The evidence detailed on the preceding pages points very directly to the conclusion that tropical cyclones are essentially convectional phenomena on a large scale. They occur in seasons and regions where high temperatures prevail; they are most effectively aided by the abundant condensation of water vapor from air at high temperatures; their circulation in every way is like that which we should expect would follow from a convectional process on a rotating earth. Yet it must be noted that the essential fact on which the belief in their convectional character should depend is not yet a matter of direct observation. It has not yet been directly shown that the temperature of the cyclonic mass is higher than that of the surrounding atmosphere at corresponding altitudes. If observations on mountain peaks should in the future show that the cyclonic mass is not warmer than the surrounding air, the convectional theory of tropical cyclones would have to be abandoned and some other theory devised to explain the phenomena.

The student should therefore hold the convectional theory in mind as being well supported by reasonable evidence, and yet as still lacking the final element of direct demonstration; he should remember the evidence that leads to the conclusion here regarded as the most probable one; he should not memorize the conclusion alone. Recognizing convection as a process characteristic of gases, easily produced by experiment on small or large scale, observable in natural processes of various dimensions, as in the "boiling" of warm air over hot sandy surfaces, in the formation of cumulus clouds, in the movement of land and sea breezes, he should appreciate the arguments that lead to belief in the convectional origin of the general circulation of the atmosphere between the equator and poles and between the continents and the oceans. He might thus, indeed, on beginning the present chapter, be prejudiced in favor of a convectional origin for tropical cyclones; yet if he would be guided by the true spirit of scientific inquiry, he must maintain an unsettled opinion as long as the evidence is incomplete or contradictory; he must adopt conclusions only where the evidence is complete and convincing; he must ever

hold his mind open to new evidence, even if it bring about the abandonment of accepted beliefs. He may, if desirable, quote the conclusions of others, and if well read he may thus become widely informed; but he will fail to gain the best benefit that comes from careful study if he does not reach opinions and conclusions for himself, forming them only as fast as the evidence that may support them is clearly understood.

The fact of the occurrence of tropical cyclones in certain regions and seasons is not doubted, for the occurrence of their violent winds and heavy rain is a matter of repeated observation. The combination of the surface winds to form a vorticular hurricane is demanded by numerous observations carefully studied; fifty years ago this was a matter for discussion, but at present it need not be questioned. The further interpretation of the surface winds and the upper currents as an inflow below and an outflow above, combined with a whirling ascent around the center, is still doubted by some, although the evidence in its favor seems conclusive to most meteorologists. But when it comes to the explanation of this interpretation as a convectional overturning, begun by the heat of the calm air in the doldrums, afterwards supported in good part by the liberation of latent heat from the condensation of water vapor, and developed into true cyclonic violence by the deflecting force of the earth's rotation, — all this is manifestly theoretical to a high degree; and belief in it is warranted only after the convincing nature of its support is appreciated.

There is a correspondence in the progress of the explanations that have been given in Chapter VI for the general circulation of the planetary winds, and in the present chapter for the occurrence of tropical cyclones, that deserves brief mention. In the first explanation, the theory of simple convectional interchange between the warm equator and the cold poles was at fault, because it involved the occurrence of high pressure at the poles, while the pressure there is really low; but on supplementing the simple convectional theory by the effect of the earth's rotation, the low pressure at the poles is perceived to be an essential feature of a convectional circulation on a rotating globe; and the theory is really strengthened by its survival of this ordeal. Again, when tropical cyclones were seen to possess a convectional inflow, ascent and outflow, initiated in the warm and moist air of the doldrums, no reason was at first perceived for their limitation to certain seasons. The doldrums are always warm and moist, and are therefore always ready to promote convectional storms. But here, again, the introduction of the omitted effect of the earth's rotation fully accounts for the special seasonal and geographical distribution of cyclones in the tropical seas; and the modified convectional theory is therefore accepted with redoubled confidence. Hence in both these explanations the simple convectional theory fails to account for certain significant phenomena — the low polar pressures in the first case; the peculiar distribution

of cyclones in the second case — and in both explanations, the introduction of the same omitted but essential consideration, namely, the fact that the movements take place on a rotating globe, completely reconciles these diverse difficulties. So similar a progression of successful theoretical explanation in two separate problems may be reasonably accepted as reacting favorably on both: indeed, the student may fairly measure his appreciation of the arguments that have been employed in these chapters by the increase of his confidence in their conclusions after their similarity is recognized.

Extra-tropical Cyclones.

232. Comparison of tropical and extra-tropical cyclones. On turning to the cyclones of the temperate latitudes, we find many features in which they resemble those of the torrid zone, and certain other features in which they

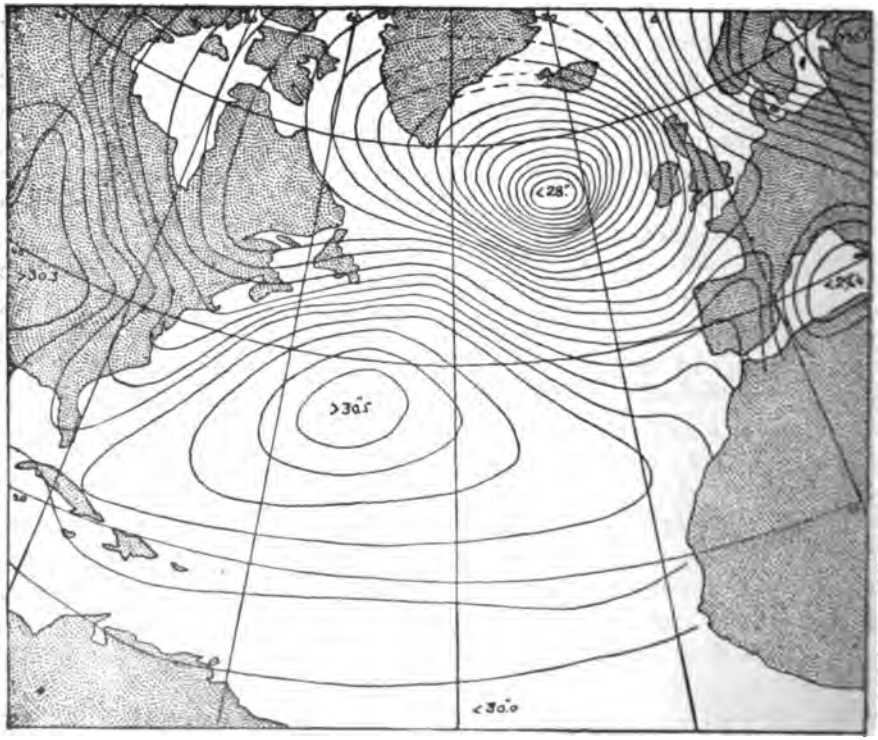

Fig. 61.

differ. Their fundamental resemblance to tropical cyclones is seen in their incurving vorticular winds, whirling to the left in this hemisphere, to the right in the other, around a moving center of low pressure. Numerous characteristic

examples of such storms are found in the magnificent Atlas of daily weather maps of the North Atlantic ocean for the year beginning August, 1882, published by the British Meteorological Council. Fig. 61 represents the isobars for every tenth of an inch around one of these storms for noon of January 14, 1883. The storm center had a pressure of less than 28.00, or an inch and a half below the normal for the place and season, as given on Chart V; it was formed by the union of several subordinate cyclonic centers, which all coalesced near the center of the North Atlantic low pressure area of winter, producing a storm of unusual severity. The contrast between the gentle and uniform gradients of the torrid zone and the strong and variable gradients of the temperate zone, as here illustrated, is very striking. Many similar examples may be found in the continuation of this Atlas by the marine observatories of Germany and Denmark, as well as in the weather maps of various countries (Sect. 325). As the surface winds obliquely approach the storm center from all sides, an upward escape must be inferred for them; and as in the case of the tropical cyclones, this is confirmed by the occurrence of extended clouds and heavy rainfall, and by the forward outflow of cirrus streamers aloft.

As with tropical cyclones, the cyclones of our latitudes vary in intensity with the depression of the barometer at the center; and here as there the greater part of the depression is to be regarded as the effect of the centrifugal forces of the revolving winds; but the greater part of these forces in a tropical cyclone arises from the true centrifugal force of the wind's rotation around the storm center, and is only in a lesser proportion due to the deflecting force of the earth's rotation; while this relation is reversed in extra-tropical cyclones, where the deflecting force is greater than the true centrifugal force of the whirl, because of the higher latitude in which these storms occur. The central region of exceptionally low pressure and very steep gradients in tropical cyclones is relatively small, because a strong centrifugal force is produced only when the winds are whirling on a short radius; the low-pressure area of our cyclones is much larger and the gradients have a tolerably strong value for some distance around the center, because the depression of the isobars depends rather on the latitude of occurrence than on the distance of the wind from the storm center; for this reason there is less concentration of violence close to the center, and the calm and clear central space or eye is seldom sharply developed, although it is not uncommon to discover a gradual weakening or failing of the winds, and sometimes even an imperfect breaking away of the clouds, as the central area passes over the observer. The form of tropical cyclones, as defined by their isobaric lines, is nearly circular. Our cyclones are as a rule less symmetrical, and their isobars are often elongated into an oval form. In the eastern United States, the longer axis of the oval trends northeast, making a trough-like depression between the high-pressure area over the tropical North

Fig. 67 illustrates several of the characteristic features of anticyclones. It represents the condition of the atmosphere at 7 A.M. over the eastern United States on January 25, 1880. The central pressures are over 30.1; the central gradients are weak and are directed outward on all sides; the air is calm or gently moving in the central region, and flows slowly outward in a right-handed spiral over the marginal area. The sky is generally clear; the hour of observation being too early for the formation of diurnal clouds, such as often arise in an anticyclonic area during the middle of the day, especially

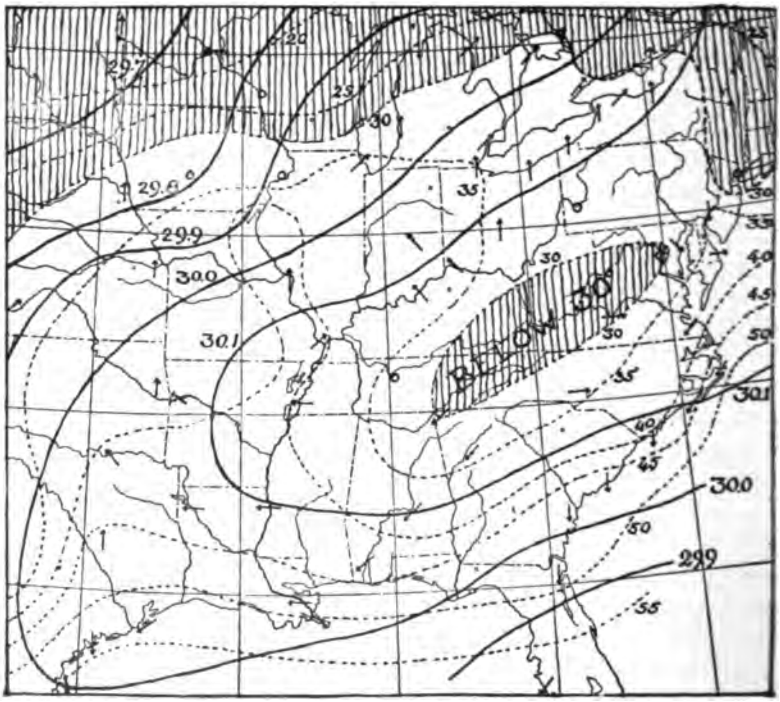

FIG. 67.

in summer. The temperature of the central clear area is lower than that of the surrounding districts, even lower than to the north; because of the free radiation from the earth through the clear anticyclonic air, while the greater cloudiness of the surrounding districts has diminished their nocturnal fall of temperature.

In winter, when insolation is brief and weak, the surface temperatures in anticyclonic areas are prevailingly low, especially on land, by reason of the rapid cooling of the ground by radiation. In summer time, they are low at night for the same reason; but they are then relatively high in the day-time, on account of the strength and long duration of the insolation that enters

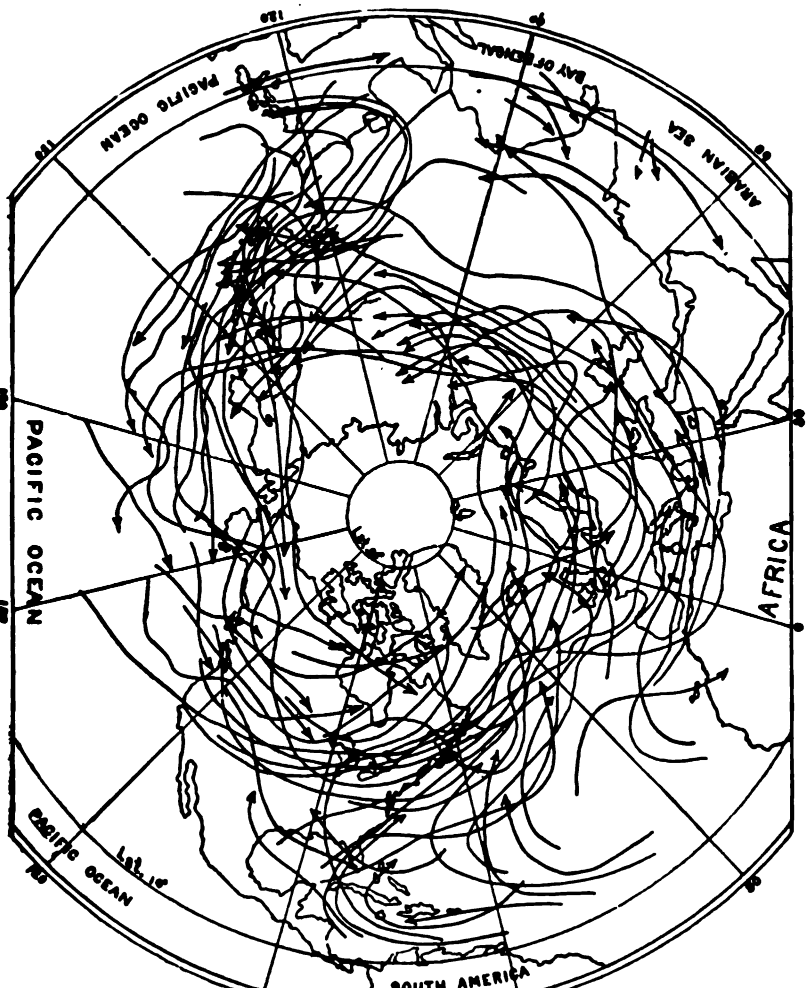

Fig. 62.

torrid zone (Sect. 240); but on entering the temperate zone and recurving towards the east, the tropical storms take on all the features of cyclones that have originated there. Storms of the two classes sometimes merge into a single center, as if their motions before union were entirely accordant.

233. Unsymmetrical form of extra-tropical cyclones. One of the strongest contrasts between the two classes of storms is found in the distribution of temperature, clouds and rainfall, with respect to the center of low

pressure. The cause of this contrast may be readily understood by comparing the surroundings of cyclones in the two zones. The oceanic area of the torrid zone is a vast region of remarkable uniformity; for hundreds of miles on all sides of a cyclonic center the temperature and humidity of the air vary but little. Inflowing currents from all sides are nearly alike as to heat and moisture; isotherms in tropical cyclones may coincide closely with isobars, and both approach a circular form. The areas of lower clouds, of

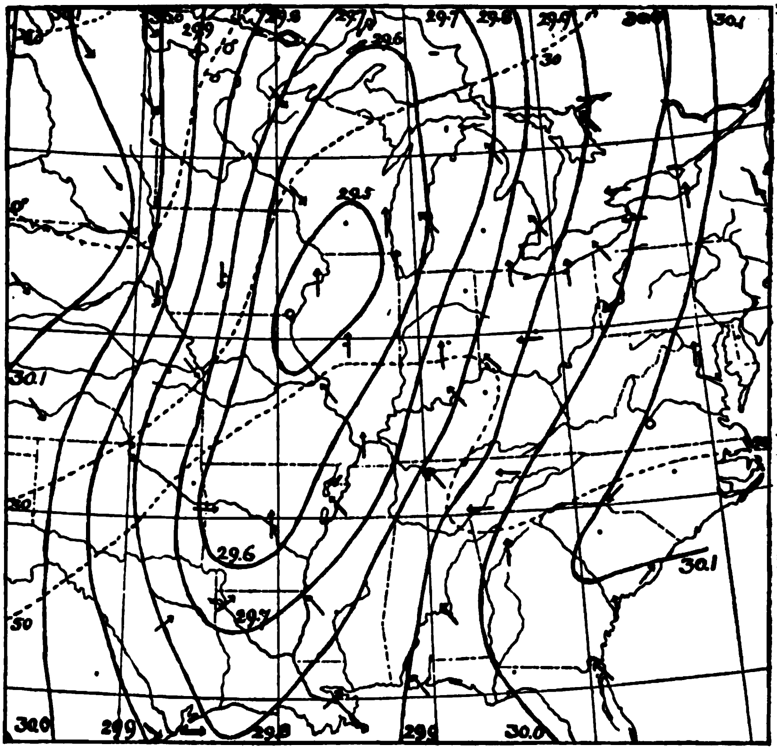

FIG. 64.

upper clouds, and of rainfall extend almost symmetrically on all sides of the storm center. Thus tropical cyclones are remarkably simple and regular in their form and in the distribution of their parts about the center.

Consider now the case of an extra-tropical cyclone moving across the Ohio valley in winter time; such a one, for example, as that of February 19, 1884 (Fig. 64). To the south and east of the cyclonic center the atmosphere is relatively mild and damp over the Gulf of Mexico and the warm waters of the Gulf Stream in the western Atlantic. To the north and west the cold, snow-covered plains of the continental interior extend for over a thousand

miles, unbroken by mountain ranges, and surmounted by a clear, cold and dry atmosphere. The inflowing southerly and easterly winds that enter the front of the storm area become cooled as they advance into higher latitudes and over the cold surface of the land, and still more as they begin their oblique ascent around the storm center; the cloud and rain areas are thus greatly extended to the south and east of the center of low pressure. The cool, dry winds from the west and northwest become for a time warmer and dryer as they advance obliquely towards the storm center, because they move over lands of higher temperature than that of their source, and because they enter latitudes where insolation is more effective in warming them; and these causes of increased heat and dryness must be overcome by the cooling of ascent, as the winds whirl around the center of low pressure, before any clouds and rain can be formed in them. The cloud and rain areas are therefore much less extended to the west than to the east of our cyclonic storms. The dissimilar temperatures at the source of these winds naturally result in a strong distortion of the isotherms within the cyclonic area; they are carried

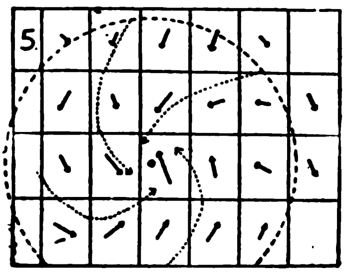

FIG. 65.

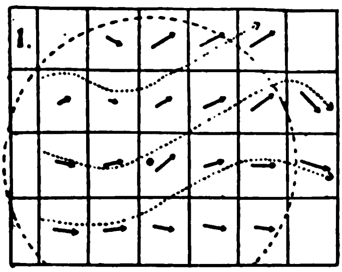

FIG. 66.

northward in front, and southward in the rear of the storm, as appears with exceptional distinctness in the illustration given above. Furthermore, the overflowing cirrus clouds, radiating in all directions somewhat eccentrically over the storm, extend their plumes much further in advance of the center than backwards from it, as might be expected from the occurrence of these disturbances in a latitude where the upper currents of the atmosphere move eastward at a high velocity. The feathery cirrus streamers are often visible a day or more before the arrival of the lower clouds. At the higher cirrus level, five or more miles above the sea, the whirling motion so apparent in the lower winds is reduced to a general eastward drift, varying but little from the velocity and direction of the higher currents of the general winds.

The results of systematic observations on the winds and cirrus clouds at Blue Hill, Mass., within the area of cyclonic storms, are presented in Figs. 65 and 66:[1] the rectangles in these diagrams being five-degree "squares" of

[1] For a fuller account of Figs. 65, 66, 68, 69, see an article by H. H. Clayton in the American Meteorological Journal, August, 1892

latitude and longitude. Fig. 65 shows the inflowing vorticular winds on the summit of the hill with respect to the cyclonic center; the irregular form of our cyclones almost disappearing in this combination of many examples. Fig. 66 shows the movement of the cirrus clouds above the cyclonic area; the deflections from the prevailing eastward movement with respect to the cyclonic center being such as to indicate a somewhat irregular outflow toward the margin of the region; the stronger outward movement to the north being attributed to the relatively high temperatures on the east of the storm.

In the frequent mention of whirling and ascent in our cyclonic winds that will be met with in succeeding paragraphs, these statements regarding the deformation of the cyclonic whirl must be borne in mind. The comparative symmetry of the winds and clouds around the vortex of tropical cyclones is not observed in our latitudes. The whirling is distinct enough in the lower winds, but the rotary motion seems to be brushed forward and obliterated at the height of the cirrus clouds. The ascensional movement about the central region is of course in no cases vertical, but always compounded with the whirling or advancing movement of the winds.

The want of symmetry is so well marked in the stronger winter cyclonic storms of our latitudes that some meteorologists are disposed to exclude them from the cyclonic class. This, however, seems to be hardly warranted; for the peculiarities are all such as would result from the occurrence of whirling storms in regions of varied, instead of uniform surroundings, and in a rapidly moving, instead of in a relatively quiet portion of the earth's atmosphere. There does not seem to be a fundamental difference in the movement of the winds in the two classes of cyclones, however different their causes may be found. The storms of the two zones not only exhibit a distinct relationship; those of the torrid zone gradually lose their symmetry when they advance into the temperate zone, and take on all the unsymmetrical features of the storms of extra-tropical latitudes. Numerous instances of this transformation may be found in the Atlases of the North Atlantic, mentioned in Section 232.

234. The center of extra-tropical cyclones. The clear central eye of tropical cyclones is not often displayed in the cyclones of our latitudes, especially on land. Our cyclonic winds decrease somewhat on the weaker gradients near the center, but the clouds do not often break away, unless the surrounding winds are of exceptional violence. It has been concluded from this that the spirally inflowing lower winds do not gain so great a centrifugal force as to prevent their being drawn into the central district of low pressure, which is formed by the more violent whirling of the winds at higher levels. In this respect the surface winds of our cyclones may be compared to the surface member of the planetary circulation, which flows obliquely towards

the low-pressure area of the polar regions; the low pressure there having been caused by the much more active whirling of the upper and middle members of the circulation (Sect. 136). Such a relation might be well expected in our cyclones on land, because the lower winds are there held down to moderate velocities by the greater resistances that oppose them; and they are therefore more subject to the control of the central low pressure as determined by the stronger winds at greater altitudes.

235. Control of weather by cyclones. It is manifest from the association of areas of cloud and rainfall with cyclonic centers, from the deformation of the isothermal lines before and behind them, and from their frequent occurrence and their generally regular and rapid advance in an eastward course, that the changes of weather in the temperate zone must be largely controlled by the passage of cyclonic storms. As they draw near, the sky becomes overclouded; the prevailing westerly wind falls away, and is succeeded by a wind from some easterly direction, faint at first, but increasing as the falling barometer and heavier clouds and rain betoken the approach of the storm center; the temperature rising, if the center passes north of the observer, until the wind veers through the south to the west, bringing a cool or cold current with rising barometer and clearing sky: the temperature remaining relatively low, and the wind backing from the east through the north to the west, if the center passes south of the observer. These changes are of great practical importance; they will be more fully considered in Sections 243 and 315: but we have first to examine the origin and movement of our cyclonic storms.

236. Non-convectional origin of extra-tropical cyclones. When we come to seek the cause of the cyclones of our latitudes, it is seen that one of their characteristics distinguishes them strongly from the cyclones of the torrid zone. This is their greater frequency and intensity in winter than in summer.

With this contrast in mind, we must inquire whether there is good reason to think that extra-tropical cyclones may, like the tropical cyclones, be regarded as convectional storms, to which a whirling motion is given by the deflecting force of the earth's rotation. This question must at present be answered most probably in the negative, in spite of the many other likenesses between the two classes of storms. When our cyclones are traced back to their beginning at one place or another, on land or water, there is not found, even in summer, any such persistent and distinct indication of instability and convection as appears in the doldrums, where the tropical cyclones begin. More than this, the occurrence of extra-tropical cyclones with increased intensity in the winter season forbids the supposition that they arise as a rule

from the convectional overturning of an unstable mass of air. If reference is made to the sections describing atmospheric instability and convection (52–54), it will be seen that the winter season is precisely the time when convection is most unlikely. In that season there is a relatively slow vertical decrease of temperature, while directly the opposite condition is necessary for instability. Moreover, in winter, when the lower air is prevailingly cold, the amount of latent heat liberated by the condensation of vapor in ascending currents of air is small compared to that liberated by an equal ascent of warmer, moister air in the summer time (Sect. 197). Latent heat, which has been shown to play so essential a part in the working of tropical cyclones, is not so effective an aid to storm action in winter as in summer; and yet it is in winter time that our cyclones possess their greatest violence. Spontaneous convectional action, therefore, does not seem to be the chief cause of extra-tropical cyclones. It may be that some of our cyclones, especially in summer and on land, are of convectional beginning; it surely is not wise in the present state of meteorology to exclude convectional action from all share in beginning and maintaining these storms; it is certain that the latent heat liberated from their rains aids their action; yet it would seem prudent to search for some other cause of their origin and action than convection; to look for some cause whose value shall be greatest in that season when the cyclones have their greatest frequency and activity.

237. Origin of extra-tropical cyclones as eddies in the circumpolar winds of the terrestrial circulation. The only cause of this kind that has yet been discovered is the general circulation of the terrestrial winds around the poles. It has already been explained that the upper, middle and lower members of this circulation move in a general eastward direction in the middle and higher latitudes, and with a considerable velocity; it has also been shown that, on account of the increased temperature gradient between the equator and poles in the winter hemisphere, it will be in that hemisphere that the general circumpolar winds possess the greatest velocity; while in the summer hemisphere, where the contrast of equatorial and polar temperatures is reduced, the circumpolar winds blow less rapidly. It has therefore been suggested that our extra-tropical cyclones are not spontaneous convectional disturbances, but secondary eddies driven by the general winds.

Reference should now be made to Section 131, in which the eastward course of the poleward overflow from the expanded air above the equator was explained. It was there stated that the eastward course of the wind was determined by a balance between the forward-acting acceleration (the component of the poleward gravitative force on the strong upper gradients not overcome by the deflecting force of the earth's rotation) and the backward-acting resist-

ances; but the character of the resistances was not then particularly inquired into. They should now be considered more carefully.

The resistances excited by the variation in the velocity or direction of the successive strata of the atmosphere must be very small, yet it has been suggested that these may be sufficient to produce undulations in adjacent currents analogous to those by which certain kinds of clouds have been explained (Sect. 203), but of much greater size. The short-circuit return of the equatorial overflow as it advances obliquely toward the pole over the converging meridians, explained in Section 141, may cause local crowding or congestion. The inequality of the poleward gradients in place and time must result in some irregularity in the winds around the poles. Under favorable conditions, these various causes may give rise to entangling motions of the upper winds, from which great eddies in the lower winds could be derived. The resistances of continents and mountains must contribute in some degree to the disorder of the general winds; yet only in a subordinate way, if we may judge by the occurrence of about as many cyclonic storms in the southern as in the northern temperate zone. In whatever way the disturbances are caused by the general winds, it is natural that they should be more frequent in the faster-moving circumpolar whirl of the middle and higher latitudes than in the slower-moving winds of the torrid zone; and that the middle and higher latitudes should witness more stormy disturbances in their winter season when their general winds are running rapidly, than in the summer time, when their general circulation is somewhat relaxed. All these considerations have in recent years turned the tide of opinion against accepting a purely convectional origin for extra-tropical cyclones, and directed it towards ascribing their origin in greater part to eddies in the general circumpolar winds.

A manifest difficulty in the way of this explanation of our cyclones is their long endurance. Many cyclones have been traced a third, a half, or even a larger share of the way around the north temperate zone; and it is difficult to understand how they could survive so long if produced as suggested above. At present, no satisfactory explanation of this difficulty can be given.

238. Anticyclones. Areas of high pressure, called anticyclones, are of as common occurrence as cyclonic areas of low pressure in extra-tropical latitudes. The two are intimately associated and usually increase or decrease in intensity together; the anticyclones alternating somewhat regularly with the cyclones in the procession of disturbances that marches around the poles. Just as the cyclones frequently merge into the larger areas of low pressure that lie over the oceans in high latitudes during the winter, so the anticyclones often combine with the areas of high pressure that lie over the northern continents in the colder months. Whatever explanation is given of one of these classes of phenomena should throw light on the origin of the other as well.

Fig. 67 illustrates several of the characteristic features of anticyclones. It represents the condition of the atmosphere at 7 A.M. over the eastern United States on January 25, 1880. The central pressures are over 30.1; the central gradients are weak and are directed outward on all sides; the air is calm or gently moving in the central region, and flows slowly outward in a right-handed spiral over the marginal area. The sky is generally clear; the hour of observation being too early for the formation of diurnal clouds, such as often arise in an anticyclonic area during the middle of the day, especially

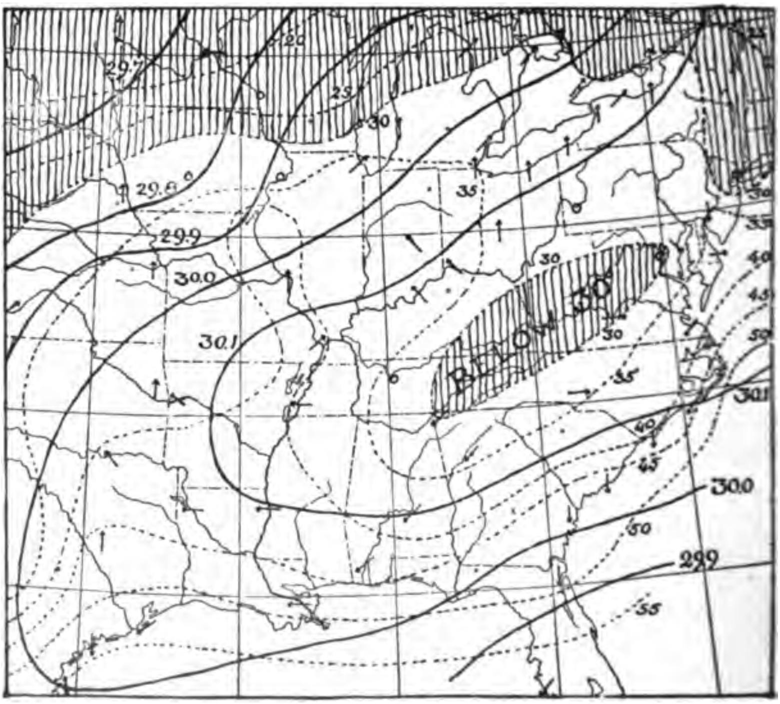

FIG. 67.

in summer. The temperature of the central clear area is lower than that of the surrounding districts, even lower than to the north; because of the free radiation from the earth through the clear anticyclonic air, while the greater cloudiness of the surrounding districts has diminished their nocturnal fall of temperature.

In winter, when insolation is brief and weak, the surface temperatures in anticyclonic areas are prevailingly low, especially on land, by reason of the rapid cooling of the ground by radiation. In summer time, they are low at night for the same reason; but they are then relatively high in the day-time, on account of the strength and long duration of the insolation that enters

through their clear sky. In both seasons, the lower air of anticyclones has a relatively large diurnal range of temperature and a distinctly marked diurnal period in the velocity of the wind.

The outflow of the lower wind from anticyclonic areas is shown in Fig. 68, which represents the average direction and velocity of the winds at Blue Hill, Mass., around an anticyclonic center. This outward movement, coupled with the prevailingly clear and dry condition of the atmosphere leads to the belief that areas of high pressure are regions of inflowing air aloft and slow down-settling about the central area. In this, as in so many other features, they are the opposite of areas of low pressure, or cyclones; hence the name, anticyclone, suggested by Galton in 1863. The occurrence of an upper inflow is confirmed by the observations of such cirrus clouds as wander near them. The average direction and velocity of cirrus cloud movement over Blue Hill within anticyclonic areas, Fig. 69, are changed from the general eastward

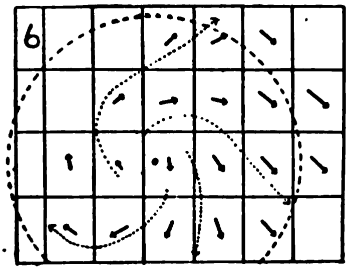

FIG. 68.

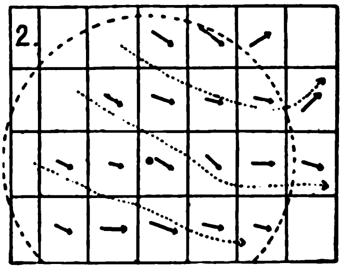

FIG. 69.

movement of the upper clouds in such a way as to indicate a somewhat irregular inward movement with respect to the advancing anticyclonic center. Further confirmation of the descent of air within anticyclones will be found in Section 249.

If anticyclones are convectional phenomena, in which the air descends spontaneously by its own weight, they must as a whole have a relatively low temperature; the converse of convectional cyclones, in which the temperature of the mass must be relatively high. According to this theory, the warmth of cyclones would be gained chiefly in the lower air, where insolation is best applied to raising the temperature of the atmosphere; and the low temperature of anticyclones would be referred to radiation from the upper air: but as this is a relatively inactive process, the cyclones would, as a rule, take the initiative, and the anticyclones would follow as consequences of cyclonic overflow aloft. The anticyclones would then be explained as the overlapping of two or more adjacent pericyclonic rings.

If, on the other hand, cyclonic and anticyclonic disturbances are produced by the irregular flow of the general winds, it is probable that these disturbances

would originate in the higher regions of the atmosphere, where the winds blow much faster than near the earth's surface. The differences of pressure produced at high altitudes would be felt down to sea-level; and as the lower winds move with comparative slowness, they would be governed by the gradients thus imposed on them by the irregular movements of the upper winds. According to this theory, an area of high pressure or anticyclone would be perceived at sea-level beneath a district where the upper currents crowd together; and an area of low pressure or a cyclonic storm would be developed beneath a region where the upper currents are somewhat divergent. Between the areas of high and low pressure thus produced there would be a constant play of the lower winds; they would run out from the centers of high pressure and their outflow would be supplied by a slow descent from aloft, and such areas would be consequently dry and free from lower clouds; the lower winds would run towards and whirl around the centers of low pressure, gathering vapor on the way, obliquely ascending there and becoming cloudy and rainy. As the crowding of the upper currents appears to be the more effective of the two processes here concerned, the anticyclones would, according to this explanation, take the initiative, and the cyclones would be regarded as relatively secondary.

239. Test of the theories of extra-tropical cyclones and anticyclones. Some impartial test of the two theories by which stormy disturbances in the general winds are explained is now needed; and an admirable one has been presented by Dr. Hann of Vienna in his studies of the temperatures prevailing in cyclones and anticyclones as recorded at the mountain observatories of the Alps. In order to appreciate the force of his arguments, we must first review the contrasted consequences of the two theories under discussion.

If cyclones and anticyclones are convectional phenomena, the former must be regions of relatively high temperature, and the latter of relatively low temperature, when compared with one another or with the surrounding atmosphere. The isobaric surfaces of the warm cyclones must be held apart, in order to produce the inward gradients below and the outward gradients above, as explained in Section 93, although the shape of the upper isobaric surfaces may be depressed by the centrifugal forces of the whirling winds, as explained for the case of tropical cyclones in Section 229. The isobaric surfaces of the cold anticyclones must be pressed closer together in order to produce the centripetal gradients aloft, where the winds flow inward, and the centrifugal gradients below, where the winds move outward. As both the cyclones and anticyclones endure for days or weeks together, it follows that the atmosphere from which their indrafts are supplied should be relatively warm and moist in its lower levels, in order to produce the instability on which these convectional disturbances are supposed to depend. In the case of the cyclones the

surface air must be warm enough, or warm and moist enough, to maintain a higher temperature than that of the surrounding air through which it rises, in spite of its cooling by expansion in ascent. In the case of the anticyclones the upper air must be cold enough to remain at a lower temperature than that of the surrounding air through which it settles down, in spite of the increase of temperature by compression during descent.

If cyclones and anticyclones are driven eddies, forced to move by the energy of the general circumpolar winds, no such instability need be assumed. The air of a driven eddy near a street corner in a blustering wind is not necessarily warmer and lighter than the air through which its whirling currents are raised; it may be heavier than its surroundings, as is the dust that it bears aloft, and owe its ascensional motion to some external force stronger than its own weight, instead of rising spontaneously like a hot desert whirlwind. In like manner, the air of an extra-tropical cyclone in the boisterous circumpolar circulation is not necessarily lighter than its surroundings; the air as well as the clouds that it sustains may be heavier than its surroundings; it may be driven up by some greater external force, instead of rising spontaneously like the air of a warm and moist tropical cyclone. So conversely with anticyclones; if they do not sink by their own weight, but are crowded down by the accumulation of higher currents above them, the heat that they acquire by compression may make them even lighter, volume for volume, than the surrounding air.

The contrasted consequences of these rival theories may be more clearly illustrated by the following figures. Fig. 70 presents the sequence of changes of temperature involved in a spontaneous convectional circulation. The relatively warm and moist lower air, with temperature *A*, cools rapidly at the ordinary adiabatic rate, *AB*, until its dew point is reached and condensation of vapor begins. Cooling is then retarded to the slower rate, *BC;* but in upper levels of the circulation additional cooling is caused by radiation, chiefly from the clouds, and by mixture with the colder upper air; hence the actual cooling of the ascending members of the cyclonic area is better represented by the line, *BD*. Now, in order that the circulation should be maintained and that the air which has ascended in cyclones should descend spontaneously in anticyclones, it must be assumed that a considerable additional cooling takes place during the passage from the upper part of the cyclonic area to the upper part of the anticyclonic area; this being indicated by the line, *DE*. As

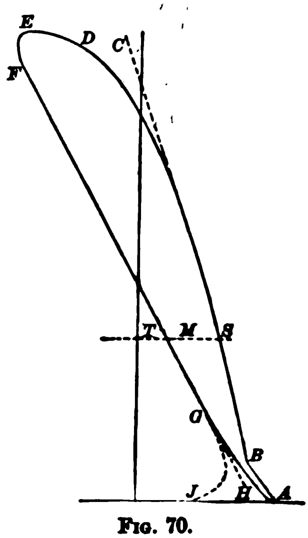

FIG. 70.

descent in the anticyclone begins, warming by compression soon overcomes cooling by radiation, and for the greater part of the descent the increase of temperature follows closely the adiabatic rate, *FH;* but on approaching the ground, departure from the adiabatic rate will be made in accordance with the temperature of the land or sea over which the descent takes place. In winter on a continent the change of temperature in the lower part of the circulation would be indicated by *GJ;* and a distinct warming and moistening of air so cold and dry as *J* must take place before it could again spontaneously enter a cyclonic ascent. It is therefore manifest that the essential conditions of the spontaneous convection here assumed involve on the whole a higher temperature in the ascending air, as *ABD*, than in the descending air, as *FGJ;* a decided cooling, *DEF*, in the upper air; and in the winter season a cooling and warming again, *GJA*, in the lower air.

Fig. 71 represents the conditions of a driven convectional circulation. The lower air cools at the rate, *AB*, at first, and at the rate, *BD*, after it becomes cloudy; the departure of *BD* from *BC* being due to radiation from the lofty clouds, as explained in the previous case. A moderate amount of cooling, *DE*, being now assumed as occurring during the passage of the upper air from the cyclonic to the anticyclonic region, the more active anticyclonic descent causes an increase of temperature at the rate, *FH*. On nearing the earth, especially in winter and over land, the temperature curve departs from the adiabatic line, *GH*, and becomes *GA*. The looped lines here drawn indicate that the greater part of the cyclonic air is cooler than the anticyclonic, but that the reverse relation obtains in the higher regions. The altitude, *K*, at which the looped lines cross, evidently depends chiefly on the amount of cooling in the upper air.

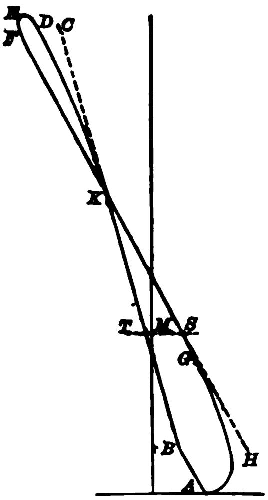

FIG. 71.

In choosing between the theories represented by the two diagrams, it seems that the latter, which assumes a moderate cooling in the upper air and a greater cooling in the lower air, is nearer to the conditions of our winter atmosphere than the first diagram. But another and more critical test has been provided by Dr. Hann's studies of records from the Alpine mountain observatories.

Let it be supposed that a mountain observatory stands at the height, *M*, and that it is frequently visited by cyclones and anticyclones. Its mean temperature, *M*, in any month will differ in one way from the mean of the temperatures observed in cyclones; and in another way from the mean of the temperatures observed in anticyclones; and the sign of these differences will

give decisive indication as to which is the better of the two theories under consideration.

The highest of the Alpine mountain observatories is on the Sonnblick, at an altitude of 10,155 feet. Although its location is not so directly in the path of frequent and strong cyclones and anticyclones as might be desired, yet its records seem to leave no doubt as to which theory they support. Dr. Hann's recent essays on this subject show clearly that as far as observations at high levels are concerned, the mass of air in anticyclones is from six to ten degrees higher than the air at corresponding altitudes in cyclones.[1]

As far as the testimony of the Alpine stations is concerned, it must therefore be accepted as being distinctly in favor of the theory that regards our cyclones as driven eddies in the general circumpolar winds; while anticyclones are regions where the upper air is forced to descend against its will. The work thus done in raising and pushing down great masses of air must be recognized as a considerable resistance which the upper winds have to overcome; and without it, they might flow faster than they do.

It should be carefully noted that the position of K, Fig. 71, is variable; that with greater cooling aloft, it stands lower; and that the lower this point stands, the more important is the spontaneous convectional action of the upper part of the entire circulation; but all of this is at present simply a matter of speculation. It should be further noted that whenever condensation takes place in a cyclone, or evaporation in an anticyclone, the work to be done in driving these disturbances is somewhat diminished. In the cyclone, condensation liberates latent heat, and the raised air is then maintained at a higher temperature, and hence in a more expanded condition than it would be otherwise; some of the uplifting is then done by the liberated energy, received from absorbed insolation long before and hundreds or thousands of miles away. It is probably for this reason that the winter cyclones that traverse the northern United States generally increase so greatly in energy as they approach the Atlantic coast, and receive an inflow of warmer and moister air from over the ocean. Conversely, when occasional clouds are evaporated in the upper part of an anticyclone, its warming in descent is retarded, and it is more easily compressed and driven down. This, however, is a small aid compared to that given by cloud-making and rainfall in the cyclones.

No test by means of mountain observatories has yet been applied to tropical cyclones; but the conditions of Fig. 70 may be reasonably interpreted as favoring their spontaneous convectional origin. It should be noticed that the initial temperature at the beginning of cyclonic ascent is higher in the first of

[1] The comparison here drawn between temperatures at the same altitude in cyclones and anticyclones should be more strictly drawn between altitudes where the same pressure occurs in the two; but if allowance be made for this, the excess of temperature is still distinctly in favor of the anticyclone, according to Dr. Hann's figures.

the two figures; this construction having been adopted in order to apply the diagram to tropical examples. The cooling during ascent, indicated by the line, *ABD*, Fig. 70, is consequently slower than that of the same line in Fig. 71; because the first example represents a beginning at high temperature and humidity, when latent heat is plentifully set free. But the chief difference between the two figures is found in the amount of cooling assumed in the upper air; and it is manifest that the greater cooling of Fig. 70 is not consistent with the conditions of the torrid zone. It may therefore be suggested that in tropical cyclones the ascending air does not soon return to sea level, but remains for a considerable time at a great height, gradually losing its abnormally high temperature there; and in confirmation of this, the absence of distinct anticyclones in the torrid zone may be adduced. The increase of temperature indicated by *FGA* must therefore be taken to apply to air masses settling down towards sea level around the cyclonic center, but not supplied by immediately previous ascent. But whatever value these diagrams possess, it must be continually borne in mind that they are for the greater part speculative; and their possible but imperfectly proved consequences must be clearly separated from consequences more closely based on observation.

Progression of Cyclones.

240. Progression of cyclones. The cyclones of both the torrid and temperate zone have been described as travelling storms. They generally advance along well-defined tracks, peculiar to the region of their occurrence. Distinction must be carefully made between the velocity of the winds around the cyclonic center and the velocity of progression of the entire storm from place to place; there is no essential relation between the two quantities. The violent hurricanes of the torrid zone move slowly along their tracks; the cyclones of the temperate latitudes are sometimes violent when advancing slowly, or of moderate strength when advancing rapidly; but the reverse relation is also observed. No general connection exists between the two velocities.

The simplest and most probable explanation of the movement of cyclones is found in the fact that they are disturbances set up in an atmosphere that is already in motion. This is plainly the case with extra-tropical cyclones; their movement accords so well with the average direction and with the seasonal changes of the general winds that there can be little doubt that they advance with the atmosphere in which they are formed. It must, however, be recognized that it is difficult to conceive of a cyclonic whirl maintaining its action in an atmosphere whose velocities in the upper and lower layers differ so greatly. The whirl cannot drift bodily; it must continually be reconstituted as it advances. Cyclonic tracks have been shown in Fig. 62. They exhibit a striking agreement with the course of the general winds around the pole.

They move faster in winter than in summer, as should be expected from the winter increase in the velocity of the circumpolar winds. Cyclones over the eastern United States move nearly twice as fast as over western Europe, as should follow from the difference in the velocity of the lofty cloud-bearing currents in the two regions (second Table of Sect. 212). The following results have been determined by Loomis.

AVERAGE VELOCITY OF PROGRESSION OF CYCLONES.

United States	28.4	miles per hour.
North Atlantic (middle latitudes) . . .	18.0	"
Europe	16.7	"
West Indies	14.7	"
Bay of Bengal and China Sea	8.5	"

MONTH.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
United States	33.8	34.2	31.5	27.5	25.5	24.4	24.6	22.6	24.7	27.6	29.9	33.4
Europe	17.4	18.0	17.5	16.2	14.7	15.8	14.2	14.0	17.3	19.0	18.6	17.9

The relation of the tracks of tropical cyclones to the course of the general winds is not at once apparent; but the general movement of these cyclones around the western border of their oceans, as illustrated in Fig. 62, suggests their control by the eddies of the general winds, explained in Section 157. Even in the doldrums, it must be remembered that only the lower air is calm. At greater heights, the equatorial overflow turns westward and towards the pole; hence the storms forming near the equator always recede from it. Their movement is at first slow and obliquely westward, somewhat accordant with the course of the trade winds; but on passing latitude 25° or 30° and entering the zone of the westerly winds, they turn obliquely to the east and quicken their pace. Cyclones that originate in the eastern part of an equatorial ocean move like the others obliquely poleward and westward, but they seldom succeed in escaping from the torrid zone against the inflow of the surface winds.

241. Effect of rainfall on progression. An additional cause has been suggested for the eastward movement of extra-tropical cyclones. The larger area of rainfall occurs on their eastern side, within the body of warmer winds that move poleward. Whatever aid is given to the action of a cyclone by the liberation of latent heat is therefore unsymmetrically placed continually to the eastward of the center, instead of symmetrically around it, as in tropical cyclones; and as far as this is an effective cause of motion, it tends to transplant the storm eastward; not by moving it along, but by continually re-creating it to the east of its previous position.

A consequence of the unsymmetrical distribution of temperature here referred to is found in the later arrival of the lowest pressures near the center of our cyclones on mountains than at lowland stations near by. The isobaric surfaces are held further apart by the relatively high temperature of the air in front of the storm; they are pressed closer together in the rear where the air is colder. As seen in vertical cross section, they would be deformed from a symmetrical to an unsymmetrical arrangement. The line[1] joining the lowest points in the successively higher isobaric curves therefore leans backwards, and when the lowest pressure is felt at sea level, the pressure is still falling at adjacent mountain stations.

242. Effect of progression on the velocity and direction of cyclonic winds. The progression of cyclones is generally spoken of as if the entire storm moved bodily forward. This seems to be true in large measure for cyclones at sea; but it is not true for the lower winds of cyclones on land.

The bodily advance of the whole storm would require that the winds on the two sides of the storm track should have different velocities, as felt by observers who do not advance with the storm. The winds on the right of the track (in the northern hemisphere), moving around the center in the direction of the storm's advance, would be increased by the velocity of progression; while the winds on the left side of the track, moving with same velocity around the center, would be decreased by the velocity of progression. Hence the winds on the two sides should, under this supposition, differ in average velocity by about twice the velocity of the storm's progress. At the same time, the inclination of the winds towards the center should be increased in the rear of the storm and diminished in the front.

Results of this kind have been found in the case of several tropical cyclones at sea; a number of West Indian hurricanes have been found to have more violent winds on the right than on the left of their tracks, and a greater inclination of the winds in the rear than in front; here the velocity of progression is small, and the velocity of the winds is high, and consequently the difference in the estimated strength of the winds on the two sides is not great. A much clearer illustration of this relation is found in the cyclones of the North Atlantic in temperate latitudes. Westerly gales are common on the south of the cyclonic centers, but easterly gales are rare on the north. As the general velocity of progression is here about fifteen miles an hour eastward, the winds on the two sides of the track might have velocities of thirty and sixty miles an hour to an observer who did not advance with the storm center, although the winds on all sides of the storm had equal velocities of forty-five miles an hour around the center.

[1] This line is sometimes called the axis of the storm. The term is misleading, because in the case of whirling on an axis, it is always implied that the motion takes place in planes at right angles to the axis; and this is not the case here.

A bodily drifting of cyclones is also apparent at moderate altitudes in the atmosphere. On Mt. Washington, for example, where regular observations were maintained for a number of years at an elevation of 6,279 feet by our Signal Service, easterly winds on the northern side of the centers of low pressure were generally faint, but the westerly winds on the southern side often attained great violence. The inclinations also showed a systematic variation. At still greater altitudes, where the general winds move very fast, the clouds show a whirling spiral outflow with respect to a rapidly-advancing cyclonic center, and yet all its parts would have an eastward motion with respect to the earth's surface, as appears to be the case in Fig. 66.

The case is quite different with the lower cyclonic winds on land. In our country, the average eastward progression of cyclones is twenty-five or thirty miles an hour. If the storms were bodily carried over the earth, the winds on the two sides should differ by fifty or sixty miles an hour; that is, there might be a calm on the northern side, while a gale of sixty miles an hour raged on the southern. This is by no means the fact. The winds do not differ greatly in velocity or inclination on the two sides of our storm centers. Their velocity depends much more on the value of the local gradients than on the velocity of the storm's progression or on their position in the storm.

It has therefore been supposed that at sea, where there is relatively little friction, the advance of a cyclonic whirl is effected by a forward carriage of the whole commotion; but on land, where the surface is much more irregular, the advance of the whirl as a whole is limited to its middle and higher parts, while its lower winds move only in accordance with the variations of pressure that are brought on them from above.

243. Veering and backing of winds caused by passage of cyclones. Records of the wind at stations in the temperate zone have long shown the occurrence of systematic but unperiodic shifts in the direction of the wind along with changes in other weather elements.

These were recognized and popularly used as means of foretelling weather changes long before any knowledge had been gained of the vorticular whirling of cyclonic winds or of the progressive motion of cyclonic storms. The alternation of northerly and southerly winds was by some ascribed to the varying success in a struggle between polar and equatorial currents; but it may be now confidently asserted that this is impossible, because such currents in temperate latitudes must flow prevailingly from the west, with relatively small north or south components (Sect. 147).

The recognition of the control of ordinary weather changes by cyclonic storms of greater or less size and intensity was first clearly made by Redfield in 1834, and constitutes an important advance in the understanding of atmospheric processes. Redfield's generalization was fully confirmed some thirty

years later, when daily weather maps were prepared in different countries, and it now serves as the most important principle used in the daily official predictions of the weather. It will be briefly explained here, and more fully illustrated in Chapter XIII.

Fig. 72 is an ideal diagram, representing the distribution of the various weather elements around a well-developed center of low pressure. The concentric oval lines are isobars for every two-tenths of an inch; the curved broken lines are isotherms for every 10° F. Imagine the entire disturbance

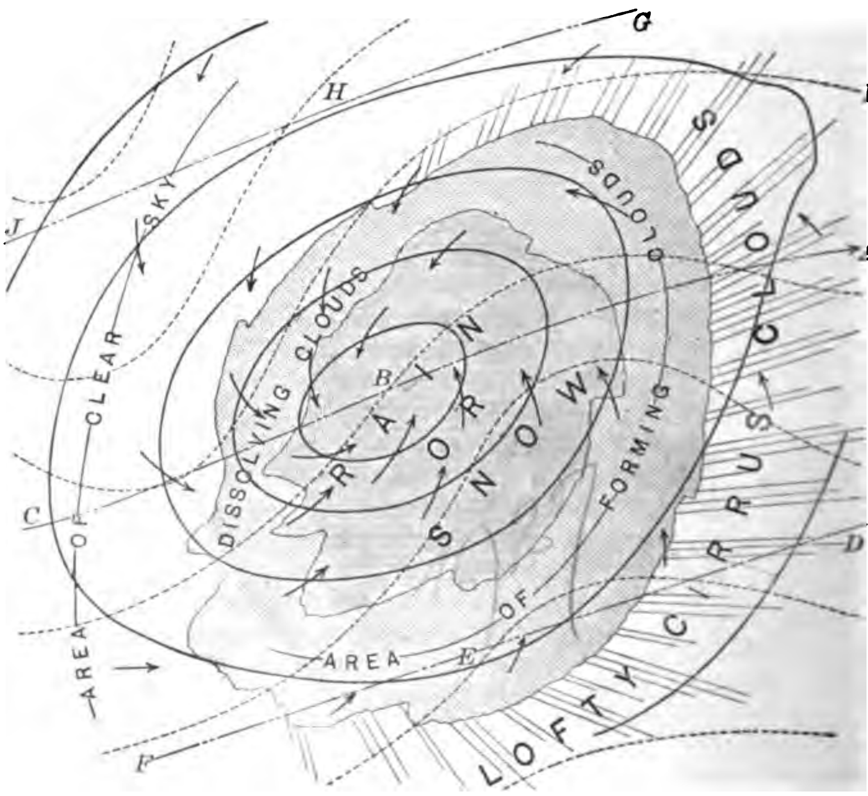

Fig. 72.

to advance in an east-northeast direction at a rate of thirty miles an hour, or 720 miles a day. An observer at any station over which it passed would stand successively under different parts of the storm, and would experience the various kinds of weather that it brings together. If his station lay on the track of the storm center, his successive positions within the storm area would fall along the line, *ABC;* if the storm center passed to

the north, his station would be found on the line, *DEF;* if the storm passed to the south, his path through it would run on the line, *GHJ.* In the case of his lying on the track of the center, he would note at first weak southeasterly winds, increasing in strength and shifting somewhat to the south, with falling pressure, rising temperature, and increasing cloudiness with rain or snow. On the close approach of the center, when the barometer reaches its lowest reading, the winds commonly weaken; and shortly afterwards turn more or less abruptly to the northwest, then increasing to a gale, as the barometer rises and the temperature falls; the rain soon ceasing, and the clouds breaking away before the wind takes its more customary direction and rate.

If the observer's station lie south of the track, there will be no abrupt change in the course of the wind, unless the form of the storm as indicated by its isobaric lines is very unsymmetrical; ordinarily the winds begin in the south or southeast, and shift with considerable regularity through the south to the west or northwest; this direction of change being called veering. If, on the other hand, the observer stands north of the track, the winds will begin in the east, and shift through the northeast and north to the northwest; this change being called backing. The abnormal direction of shift implied in the term "backing" is simply explained: European stations, whence these terms have come to us, lie generally to the southeast of the tracks of their stronger winter storms; hence the ordinary shift of the winds is from the east through the south to the west, or "with the sun," as it is often called; it is only when an exceptional storm center passes further south than usual, or only at the distant stations in northern Europe, that the winds turn "against the sun"; and this change consequently came to be looked on as the reverse of the normal order. While it is truly of unusual occurrence in those regions where storm centers generally pass to the north, it is manifestly the normal order of change for regions whose storms habitually pass to the south. As our cyclonic centers generally pass over the Great Lakes and down the St. Lawrence valley, observers in the United States are accustomed to veering winds; but observers in northern Canada may normally have backing winds.

The changes of wind and weather thus described may be faintly marked during the passage of cyclonic areas of slight barometric depression; but during an ordinary season an attentive observer may detect twenty or thirty unmistakable examples of cyclonic shifts. If the cyclonic center pass far to one side of the observer, its effects are hardly noticed; but if its passage be nearly central, the changes in its wind, temperature and sky will be most emphatic. If the center advances gradually, the changes in the weather elements will be slow; if the center progresses rapidly, the changes will be quickly run. If the center turn to the north or south of the ordinary path, the shifts will be somewhat abnormal and unexpected; but as most cyclones follow a tolerably regular track across our northern states, the ordinary sequence of

although there may still be a decrease of humidity; but in winter, both the heat and the dryness of the foehn are remarkable. The snow on the mountain

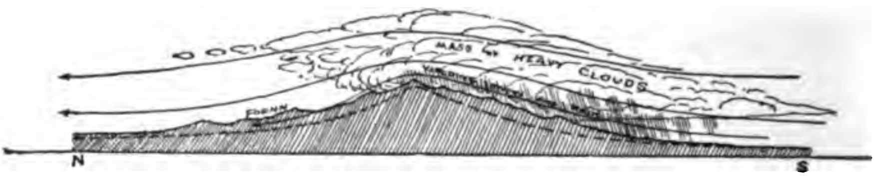

FIG. 84.

slopes disappears under its heated breath; hence the name *Schneefresser*, or snow-eater, sometimes locally given to it. Extensive fires among the wooden houses of the Swiss villages have happened at such times, the last one of the kind being that which destroyed Meiringen in the winter of 1891–92. The peculiar heat and dryness of the foehn are soon lost as it advances across the Piedmont plateau, cooling and absorbing vapor on its way.

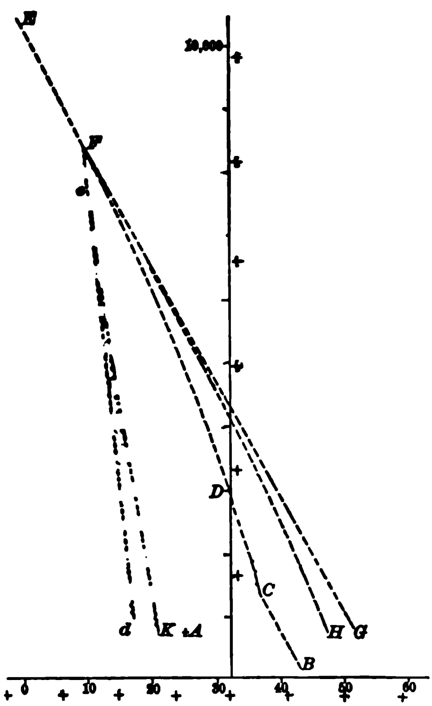

FIG. 85.

After the initiation of the foehn as thus described, it may be continued by a further supply of air coming from the plains of northern Italy, and then an additional cause for the heat and dryness of the wind is introduced. When the air is drawn away from the summit of the Alps, other air rises from the Italian lowlands, passes over the range, and descends on the leeward slopes, Fig. 84. Before ascent, the temperature of the air on the Italian plains may be represented by *B*, Fig. 85, while the temperature is *A* in the northern valleys. During the ascent of the Italian air, its temperature falls, its dew-point is reached, the whole mass becomes cloudy, and rain or snow falls from the clouds on the Italian slope. In the first part of the ascent, before the cloud level is reached, the temperature of the ascending air decreases at the rapid rate *BC*;[1] but when clouds are formed at the height *C* and above, further

[1] It must be borne in mind that the horizontal scale of Fig. 85 indicates temperature only, and hence that its oblique lines represent only rates of cooling with ascent or descent: they have nothing to do with the inclined path of the air over the mountains, illustrated in Fig. 84, but only with the effect of the vertical components of motion during the passage.

e and *f*, Fig. 10, are examples of temperature changes under warm cyclonic winds in New England. A wind of this kind is generally damp and cloudy with rain in winter, from having cooled on its way until its upper portion at least has been chilled below the dew-point. Observations on Mt. Washington give many examples in front of approaching cyclonic centers, where the temperature in winter a mile above sea level is actually higher than that on the cold lowlands over which the wind blows; this is because the wind moves faster aloft and therefore comes more quickly from a farther source; it thus starts with a high temperature and soon becoming cloudy retains much of its heat on the way; while the wind that blows from the south closer to the earth's surface comes slowly from a less distance and is much more cooled in its northern progress. The vertical temperature gradient at such a time may be represented roughly in Fig. 73;[1] *AB* being the normal value for the place and season, and *CDEF* being the temporary value during the blowing of the southerly wind. The greatest increase of temperature is at some considerable altitude, as at *E*; and there may be thus produced an inversion of temperature for a certain distance above the earth's surface, as *CD*. Inversions of this kind, caused by cyclonic importation of little cooled air aloft and comparatively independent of diurnal changes, should be compared with the inversions at night described in Section 43, produced by the local cooling of the quiet lower air by radiation and conduction. As the reversed gradient, *CD*, indicates great stability in the lower air, winds of this class possess relatively constant strength night and day, and they are remarkably free from the day-time flurries and flaws of our northwest winds. The occasional silver thaws and ice storms of our winters (Sect. 287) depend on inversions of this kind under easterly winds; the temperature at *D* being a little above freezing, while at *C* it is below freezing.

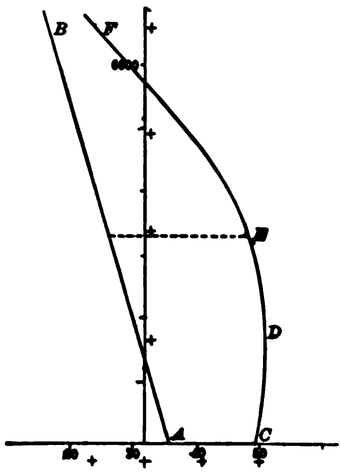

Fig. 73.

In the spring and summer, the southerly or southwesterly wind in front of cyclones in this country may be dry as well as warm, because the lands over which it then blows already possess a relatively high temperature and the strength and length of insolation in the higher latitudes do not permit the easy cooling of the wind. Inversions of temperature do not then prevail. The enervating days in the warm spells of April and May are found under

[1] All the figures of vertical temperature gradients in this chapter are largely hypothetical: they represent graphically certain probable conditions that cannot easily be described in words.

such conditions (see curve *a*, Fig. 10); the prolonged spells of intense summer heat are caused in the same way. The "hot winds" of Kansas and Texas seem to be intensified examples of this kind of wind. At all seasons, it is generally in the presence of these southerly cyclonic winds that the highest temperatures of our months are found.

It is manifest that winds of this kind must occur wherever the spiral cyclonic inflow brings warm winds over a cool region. In western Europe, although the southerly or southwesterly wind is warm, it does not produce the strong rise of temperature that accompanies it in the eastern United States, because of the much slower change of mean temperature with latitude there than with us (Sect. 82). Along the northern shores of the Mediterranean, however, when cyclonic storms pass near by, a warm wind blows northward in front of them from Africa. It is sometimes parching hot and dry, bearing dust from the desert and seriously injuring vegetation; its heat being felt almost as severely by night as by day. Such is the sirocco of southern Italy and Greece, illustrated in Fig. 81; but further north, after the wind has blown over a greater breadth of water surface, gaining vapor on the way and at the same time cooling somewhat on its northward journey towards the cyclonic center, it becomes moist and cloudy, its high temperature then making the air extremely sultry and oppressive. In Spain, the dry sirocco is called the leveche;[1] on the Madeira islands it is called the leste. The harmattan, a hot, dusty east wind on the west coast of the Sahara, may be of similar cyclonic origin.[2] In Egypt, the representative of the sirocco is known as the khamsin, from its relatively frequent occurrence during a period of fifty days in early spring, when the southward retreat of the tropical belt of high pressures allows cyclonic storms to extend their influence over northern Africa. The association of these various winds with cyclonic storms is not demonstrated in all cases, but it appears to hold true as far as they have been studied. Hence the advisability of recognizing them as belonging to a distinct class of atmospheric phenomena, deserving a particular name. The Italian name, *sirocco*, might be adopted in any part of the world, when the peculiar features of the wind are well developed. It is not intended that every light wind in front of a cyclonic center should be called a sirocco; but when the wind is active and its temperature is decidedly above the normal, this name may be properly given to it.

In the southern hemisphere, winds of the sirocco class come from the north. Such are the "brickfielders" or hot north winds of southern Australia, and the zonda of the Argentine pampas.

[1] The solano of the east coast of Spain is a cloudy, rain-bringing east wind; probably a cyclonic indraft, but not to be confused with the hot and dusty leveche.

[2] The harmattan of the Gulf of Guinea is a cool, dry, northerly wind of the winter season, like an intensified trade wind.

The "hot winds" of Texas and Kansas, above referred to, seem to possess certain special features in addition to those of the normal sirocco. The greatest heat, 105 or more degrees, is felt in narrow currents, ranging from 100 feet to half a mile or more in width, with intermediate belts of considerable breadth at less insufferable temperatures. The intense heat and extreme dryness of these winds make them very injurious to all crops. Their heat does not seem to be due simply to importation by a horizontal flow, and it has therefore been suggested that they are supplied by descending currents, warmed adiabatically. The apparent difficulty of this suggestion lies in the excessive temperature that the currents gain at the level of the ground, in virtue of which they should be lighter and not heavier than the lower air into which they are assumed to descend: but as descent from a great height must be inferred in order to produce their excessive temperature, a descent by momentum below the level of equilibrium may also be inferred; and thus the local and temporary quality of the hottest currents might be explained. Hot winds of a similar quality occur on the plains of India in the early summer; and it is possible that certain of the suffocating simooms of the Arabian and African deserts, excessively hot but free from dust, should be explained in this way; while the simooms that bear clouds of sand and dust should be associated with thunder-squalls (Sect. 255).

246. The cold wave. The warm sirocco in front of a cyclonic storm is in distinct contrast with the cool or cold equatorward wind in the rear. In summer, the latter wind possesses only moderate strength, producing an agreeable cooling after the excessive heat of the sirocco that it drives away; it may then be called a cool wave; clear, dry and refreshing, after the sultry and oppressive air that preceded it. In winter, it may be a strong wind, coming quickly and causing a rapid fall of temperature. This fall is called a *cold wave* by our Weather Bureau,[1] and the term may be applied to the wind that brings it and extended to other winds of the same kind.

The cold wave is remarkably well developed in the winter storms of the central and eastern parts of our country. When supplied by an area of strong high pressure in the northwest, it sweeps down from the cold plains of farther Canada and brings with it the low temperature of that bleak region. Its movement obliquely towards the cyclonic center that it follows is accelerated by the winter high pressure characteristic of its source in the continental center (see foot-note, page 92), and it is nowhere impeded by transverse mountain ranges. Near the track of a cyclonic storm, the cold wave arrives suddenly and in almost fully-developed strength, displacing the antecedent

[1] Technically, a "cold wave" is a fall of temperature lower than 32° in the northwest, or lower than 40° in the south, with a change of at least 20° in twenty-four hours, and without regard to the velocity or direction of the wind.

sirocco in a few hours and causing an abrupt fall of temperature, as in Fig. 10*c*; and then gradually falling away, being in this unlike the sirocco, which begins gently and only gradually acquires its full development. All the Mississippi basin and the Atlantic states as far south as Florida may feel the blast of the cold wave as it comes sweeping down from the northwest; a clear,

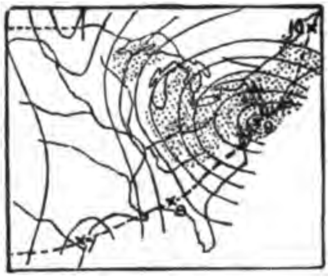

FIG. 74. JANUARY 9.

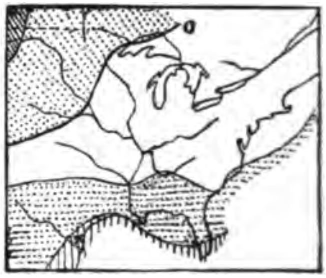

FIG. 75. JANUARY 7.

FIG. 76. JANUARY 8.

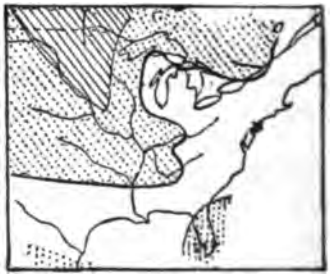

FIG. 77. JANUARY 9.

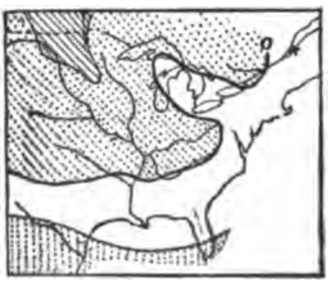

FIG. 78. JANUARY 10.

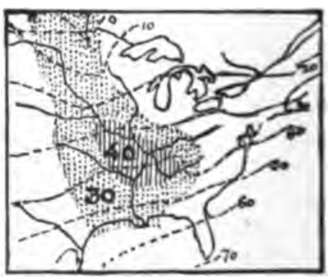

FIG. 79. JANUARY 9.

cold, dry wind, freezing the ground over which it advances, and thereby warming somewhat in its own lower layers. It may cause a continuous fall of temperature during noon-day, as is illustrated in Fig. 10*d*.

A very strong cold wave occurred early in January, 1886, as illustrated in the accompanying figures, 74–79. The cyclone behind which the cold winds

were drawn down from their source in the far northwest, lay centrally on the coast of Texas on the morning of January 7; advanced to southern Alabama on January 8, and to New Jersey on January 9, when its nearly circular isobars were remarkably well developed, as in Fig. 74. On the morning of January 10, the center lay in the Gulf of St. Lawrence. Figs. 75 to 78 indicate the spreading of the cold northerly air over the country on the four dates above named; the white space including temperatures between zero and freezing; while the lined shadings represent temperatures lower than $-30°$ in the northwest and over $+50°$ in the south. The cold air that lay in the northwest on January 7 had reached Texas the next morning, when the isotherm of zero ran nearly north and south in the middle Mississippi valley. On the two following days, the zero winds advanced up the Ohio valley, and carried temperatures below freezing even into Florida, where the orange groves were seriously injured. The retarded advance of the cold around the Great Lakes was in part due to their conservative influence, but in part also to the derivation of their winds more from the northeast than from the northwest. Fig. 79 exhibits the normal temperatures for January in oblique dotted lines; its shaded areas show the negative departures from the normals on the morning of January 9; amounting to 40° in the central Mississippi valley.

The probable value of the vertical temperature gradient in a cold wave may be represented in Fig. 80; the greatest cooling from the normal being at an elevation, *D*, of half a mile or more, where the winds are strongest. The departures of the gradient lines so greatly from the mean value in the cold wave and in the sirocco should be compared with the smaller departures illustrated in Figure 3, page 27. The latter result from changes in the temperature of relatively quiet air by the slow processes of absorption and radiation; the former result from the active cyclonic importation of new bodies of air, which bring with them the temperatures characteristic of their source, thus strongly affecting the atmosphere up to heights of more than a mile. Variations of absolute humidity are produced in the same way. Nonperiodic changes of temperature and humidity in the higher layers of the atmosphere are therefore to be regarded as chiefly caused by the changes of cyclonic winds.

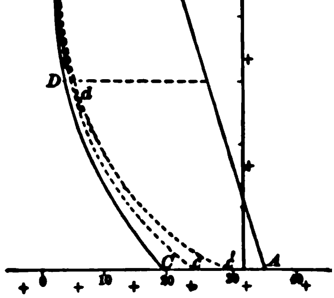

FIG. 80.

Near the earth's surface, the air of the cold wave generally becomes warmer as it advances, and its gradient curve may be represented by *dc* or *dc'*. This is particularly the case in the day-time

and in the early spring when there is no snow on the ground. The lower layers of the wind may thus become unstable, and hence there is at such times a strong variation in velocity between day and night; the wind of day-time is continually hastened by rushing flaws and flurries, as its currents roll over and over in their convectional turning, while the lower air at night is comparatively quiet. But when the cold wave blows at night over a snow-covered region, the temperature of the air near the earth's surface may decrease as it advances; the vertical temperature gradient may then be reversed in the lower air.

When blowing at high velocities and bearing a blinding cloud of snow with the temperature below freezing, the cold wave is called a blizzard. The norther of Texas and the Gulf of Mexico includes both the cold wave of winter and the cool wave of summer: this wind is frequently developed in the autumn on the western side of a tropical cyclone as it advances slowly along the recurving part of its path south of Louisiana. When following a cyclonic center in winter, the norther may cause a fall of 30° in an hour, or of 50° in two hours.

The winter cold wave of Europe is much less pronounced than with us, and comes from the northeast instead of from the northwest. The mild waters of the North Atlantic lie northwest of Europe, hence no strong fall of temperature is brought by an inflow of winter winds from that direction over Great Britain and France in the rear of a cyclonic storm that passes eastward over the North Sea toward Russia; but when the storm center moves from the Atlantic across southern France towards Italy, and at the same time an anticyclone lies over northern Russia, then western and central Europe experience a cold northeast wind, akin to our cold waves. The wind on the north of the cyclone moves southwestward from the cold northern continental area and floods Germany and France and even Great Britain with air of unusually low temperature. The severe frosts of the winter of 1890–91 in Great Britain accompanied a period of frequent northeast winds of this character.

If a cyclonic center passes far enough south to draw the cold air after it from the low plateau of central France, the wind is called the *mistral* as it flows down the valley of the Rhone to the Mediterranean; the name being derived from *magistral* meaning master. When France is snow covered under a clear sky, the air on the plateau becomes cold from local radiation, and the gentle slope of the country towards the Mediterranean aids the baric gradients in hurrying the aerial drainage and intensifying the wind; thus repeating on a smaller scale the conditions of our western plains. Marseilles may have a cold mistral while southern Italy suffers under an oppressive sirocco; as illustrated in the map of a cyclone, Fig. 81, which passed northeast across Italy on February 25, 1879; the value of the isobars being in millimeters, the

temperature in Fahrenheit degrees: the heavy wind arrows indicating stations of observation, while the fainter arrows are added to generalize the course of the wind around the center of low pressure. The distortion of the isotherms by the winds is perfectly apparent.

Central Europe never feels the excessive cold that is produced by the cold waves of the upper Mississippi valley, where temperatures of twenty or thirty degrees below zero are recorded; but the more severe winter cold of Europe is generally experienced under such conditions as have been just described.

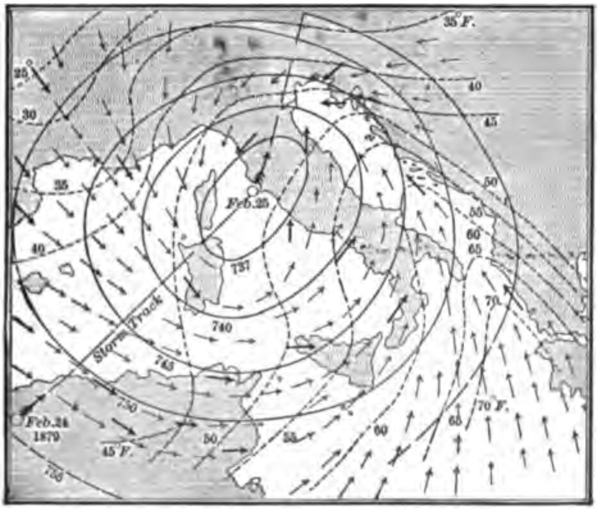

FIG. 81.

The name, cold wave, is not employed there, although it is perfectly applicable. Further east, in Russia and Siberia, where the continental extension allows a more severe winter, the colder cyclonic wind is more like our cold wave; when blowing violently and raising a cloud of fine dry snow, it is called a *buran* or *purga*, corresponding to the blizzard with us.

The southern hemisphere has cool waves from the south in the rear of its cyclonic storms; but in the absence of large land areas in high latitudes, the fall of temperature is never as violent as with us; no strong cold waves occur there. The wind of this kind in the Argentine Republic is called the *pampero*. The "southerly burster" of New Zealand also seems to belong here.

247. The bora. It occasionally happens that the cold wind drawn down from high plateaus towards a cyclonic center takes on especial features which entitle it to a separate description. The strong radiation from an elevated plateau during short winter days of weak sunshine and long clear winter

nights reduces the air that lies upon it to an abnormally low temperature. If the vertical temperature gradient over the surrounding lowlands is *AB*, Fig. 82, it may be *CD* over the plateau. Then if the air flows off of the plateau and descends rapidly to the surrounding lower ground, it is warmed by compression at the rate *CE* during descent towards sea level, and when it arrives on the adjacent lowlands it still has an unusually low temperature, *E*. Under ordinary conditions the drainage of the plateau is relatively slow, and its descent is not strong enough or in large enough volume to be particularly noticeable; but if an advancing cyclone brings pressure gradients over the plateau area in such a way as to drive the air rapidly from it, it flows down over the adjacent lowlands in large volume with great velocity; being induced to move, not only by the cyclonic gradients, but also by its own instability or top-heaviness. It is then felt as an icy blast, generally producing boisterous squalls, sometimes with snow flurries; the latter probably due to some reaction between the descending air and that which it displaces.

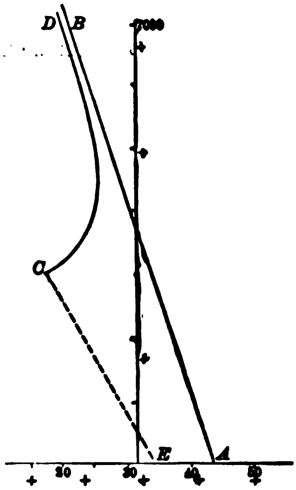

FIG. 82.

Winds of this kind are not of common occurrence. They are recognized at the head of the Adriatic sea, where the name, *bora*, is a survival of the classical Boreas, the north wind. Here the bora comes from the Istrian and Dalmatian highlands on the northeast. The mistral has already been described as owing part of its cold and violence to its descent from the low plateau of central France; but in that case, the descent is so gradual as not to warrant its being classified here. Bora winds should be looked for in our western plateau regions, where pronounced examples of their occurrence may be expected. Their recognition will constitute one of the advances that should be made in physical meteorology by local observers.

In order to perceive the curious contrast between winds of the bora type and those of the class next described, it must be borne in mind the bora requires an extended area of elevated country, where the air may be abnormally cooled, and whence it is hastily withdrawn to lower levels by cyclonic aid.

248. The foehn or chinook. One of the most peculiar members of the class of cyclonic winds is found where the indraft is required to pass down from or over a mountain range in its course towards the cyclonic center. Winds of this kind often have a strong development in the northern valleys of the Alps, where their remarkable heat and dryness were first noted and

investigated; the local name, *foehn*, there employed, is now extended to other regions as well. The occurrence of the Alpine foehn may be described as follows: — When a cyclonic storm advances from the Atlantic over central or northern Germany, the air in front and to the south of it is successively drawn in obliquely towards the center, first from middle Germany and northern Bavaria, then from the sloping plateau at the northern base of the Alps, next from the valleys among the mountains out of which the air flows to take the place of that which has moved away from the piedmont plateau; later, by the still further backward propagation of the disturbance, even the air from the mountain tops is drawn down into the valleys, and finally a supply of air for the mountain tops is derived from the further side of the range, either at the average height of the crest line or from the lowlands. The peculiarity of this wind as it descends into the northern valleys depends on the changes of temperature produced in consequence of its motion having a vertical component; while the temperatures characterizing the sirocco and the cold wave depend essentially on their horizontal motion.

When the upper air is drawn quickly down from the Alpine crests into the valleys below, it is heated by compression at the normal adiabatic rate, and from having been a cold wind on the mountain tops, it reaches the valleys abnormally warm and dry. This is particularly the case in winter; the air in the Swiss valleys *before* the approach of a cyclonic storm becomes especially cold and damp or foggy by the accumulation of cooled air that creeps down the mountain sides and settles in the depressions; while at the same time the air bathing the peaks is comparatively dry, and is relatively little cooled, and departs less from the mean of the year. Under such conditions, the vertical temperature gradient *CD* or *C'D*, Fig. 83, would be weaker than its annual value, *AB*. Now if the cold bottom air is rapidly withdrawn to supply a cyclonic indraft, the air from the level of the mountain tops must as rapidly descend into the valleys. As it descends it warms rapidly, almost at the adiabatic rate, *DJ*, while its dew-point rises but little faster than at the rate *ed*, due to compression (see page 163); and on its arrival at the valley bottom, with temperature *E* and dew-point *F*, it is much higher in temperature and much lower in humidity than was the air which occupied the valley before. In summer, these changes are not so marked, because then the air in the valley may be unduly warm, and when the upper air is drawn down to take its place, no significant change of temperature will be introduced,

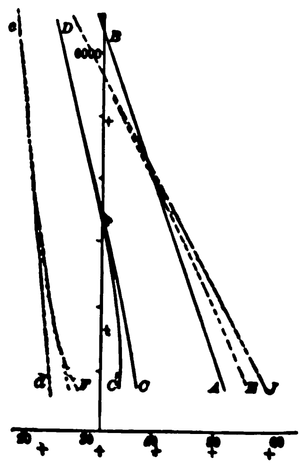

FIG. 83.

although there may still be a decrease of humidity; but in winter, both the heat and the dryness of the foehn are remarkable. The snow on the mountain

FIG. 84.

slopes disappears under its heated breath; hence the name *Schneefresser*, or snow-eater, sometimes locally given to it. Extensive fires among the wooden houses of the Swiss villages have happened at such times, the last one of the kind being that which destroyed Meiringen in the winter of 1891–92. The peculiar heat and dryness of the foehn are soon lost as it advances across the Piedmont plateau, cooling and absorbing vapor on its way.

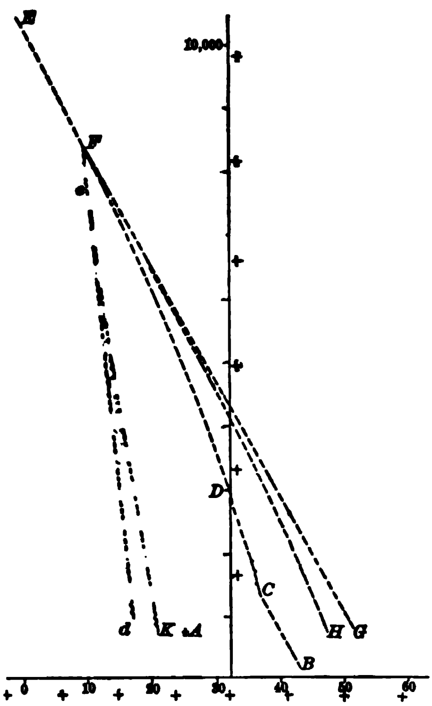

FIG. 85.

After the initiation of the foehn as thus described, it may be continued by a further supply of air coming from the plains of northern Italy, and then an additional cause for the heat and dryness of the wind is introduced. When the air is drawn away from the summit of the Alps, other air rises from the Italian lowlands, passes over the range, and descends on the leeward slopes, Fig. 84. Before ascent, the temperature of the air on the Italian plains may be represented by *B*, Fig. 85, while the temperature is *A* in the northern valleys. During the ascent of the Italian air, its temperature falls, its dew-point is reached, the whole mass becomes cloudy, and rain or snow falls from the clouds on the Italian slope. In the first part of the ascent, before the cloud level is reached, the temperature of the ascending air decreases at the rapid rate BC;[1] but when clouds are formed at the height C and above, further

[1] It must be borne in mind that the horizontal scale of Fig. 85 indicates temperature only, and hence that its oblique lines represent only rates of cooling with ascent or descent: they have nothing to do with the inclined path of the air over the mountains, illustrated in Fig. 84, but only with the effect of the vertical components of motion during the passage.

cooling is retarded by the liberation of latent heat from the condensing vapor, the rate of cooling then being *CE*, a brief ascent without cooling occurring at the temperature of freezing, *D* (Sect. 198). The temperature to which the air is reduced on reaching the level of the mountain crests is therefore not so low as it would have been if it had risen to that height without becoming cloudy.

As the wind flows down from the peaks and passes of the Alps, the clouds that it carries along are soon dissolved by the increase of temperature produced as the air descends to the northern valleys; for a great part of the vapor has been taken from the wind to fall as rain or snow on the Italian slope; the remaining cloud mass stands on the mountain crests, and is locally known as the foehn wall. As long as any cloud remains to be dissolved, the increase of temperature in the descending air goes on at the slow rate, *EF*, just as the decrease of temperature was slow during the cloudy ascent; but as soon as the cloud disappears, as at the altitude *F*, the further descent is accompanied by a rapid increase of temperature almost at the normal adiabatic rate, *FG*. On reaching the lower valleys, a temperature *H* is attained, which is greatly in excess of that of the air, *A*, in the northern valleys before the foehn began to blow, and a number of degrees higher than the temperature of the Italian air when its ascent began on the further side of the mountains. At the same time, the air will have become extremely dry; from being saturated at the height *F*, its dew-point comes to be *HK* degrees below its temperature in the valley bottom.

The increase of temperature produced by this peculiar reaction of latent heat will be stronger if the air is damp and relatively warm before beginning the ascent of the mountains, and under such conditions this second cause of the heat and dryness of the foehn may be as effective as the first; but it must be remembered that in the case of several foehn winds studied in Switzerland, the simple descent of the upper winter air is the first cause of the heat and dryness of the wind, and the liberation of latent heat is only a later and secondary cause; there sometimes being no rainfall on the southern slopes until a day or more after the fully-developed foehn is felt in the northern valleys.

The heat and dryness of the foehn are so unlike the cold and dampness of the wind on the mountain passes that it was only natural for earlier observers to ascribe the origin of the warm wind to some warm source; and it was consequently referred to the hot Sahara and regarded as an extension of the sirocco of southern Italy. This has been completely disproved. Espy and Dové were among the earlier meteorologists who suggested that the changes of temperature in vertical currents and the liberation of the latent heat from condensing vapor must be considered in its explanation; and this has been fully confirmed by modern students; especially by Hann of Vienna, to whom the suggestion of the initial cause of the heat and dryness of the foehn is due, and whose studies

have given careful demonstration of the more general suggestions of others. When thus explained, it is manifest that the foehn need not be limited to a south wind in the northern valleys of the Alps; it might occur under fitting conditions as a north wind in the southern valleys of those mountains, and such has proved to be the case; it might occur at the base of other mountain ranges, as has been abundantly shown. Wherever a lively cyclonic indraft draws the air away from the valleys at the foot of lofty mountains, leaving them to be filled by the descent of air from the altitude of the mountain crests and later by the passage of air over the range, a foehn-like wind may be expected.

These conditions are admirably developed along the eastern base of our Rocky Mountains in Montana, Wyoming, and Colorado, as well as in the Northwest Territories of Canada. It frequently happens in the winter season that as a cyclonic center moves eastward from British Columbia to Manitoba, while an anticyclone follows across Utah, an extended cyclonic circulation is developed over the mountain region. The winds that blow northward along the plains, as the cyclone advances, are soon supplanted by westerly winds; and as these descend from the mountains and flow out upon the plains, all the features of the Swiss foehn are developed. The warm and dry wind thus produced over a belt of country along the foot of the mountains is called the *chinook*. While the west wind may be damp and chilly under heavy clouds with a plentiful fall of rain or snow on the western side of the Front Range, the sky is fair or clear over the plains, and the clouds are left behind at the summit of the mountains; and although the velocity of the wind east of the mountains may be high, and its arrival may be at night, its temperature is mild or even warm, in marked contrast with the colder air that it has displaced and with the cold wave that usually follows a few days later. The warm chinook may arrive at night as well as by day (see Fig. 10, curve *g*); it quickly melts or dries up the snows of preceding storms, thus laying bare the northern plains and enabling cattle to survive the winter without protection. An isolated area of relatively high temperature may thus be produced by an extended chinook along the eastern base of the Rocky Mountains. Owen's Valley, below the Sierra Nevada in eastern California, should possess well-developed chinook winds. It lies to leeward of a lofty mountain range whose western slope is visited by heavy snow storms in the winter season; but as yet no accounts of such winds have been received from that nearly desert region.

One of the most remarkable examples of the foehn is found on the western coast of Greenland, where the cyclonic wind descending from the icy plateau of the interior becomes unduly warm and dry, raising the temperature at sea level even thirty or forty degrees above the winter mean. One of the best examples of its occurrence was during nine consecutive days in November and

December, 1875, when it was as warm in western Greenland as in northern Italy, and warmer than in Canada, Iceland, Great Britain, and over the intervening Atlantic; at one station of observation it was warmer in the darkness of the polar night than in France at noon-day.

The southern hemisphere has well-marked foehns in New Zealand, where they descend from the mountains and blow as "northwesters" across the Canterbury plains; and again at the eastern base of the Andes in the Argentine Republic, where their occurrence presumably on the equatorward side of passing cyclones is symmetrical with that of the chinook in our western states.

249. The anticyclonic calm. Although it is not at first apparent why calm air should be associated with a group of winds, still under a natural classification it seems best to place anticyclones, in which the winds are very gentle or at rest, in close association with the more boisterous cyclonic movements of the atmosphere just considered. Anticyclones possess a gentle descending movement of the air, and a quiet marginal outflow near the earth's surface, as has been explained in Section 238. They are in consequence prevailingly clear or fair and dry atmospheric regions; and in this respect they are strongly unlike the cloudy and wet air of cyclonic areas. But as their movements are of a definite kind, so are their various other features; and hence they deserve consideration as phenomena having persistent and recognizable characteristics. Being prevailingly clear, their surface air takes on the features of their season, and they must therefore be considered under different heads for winter and summer.

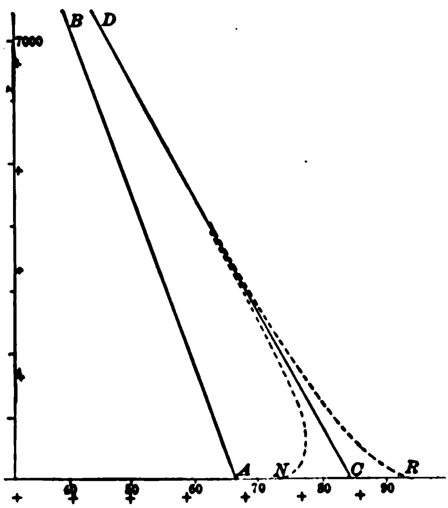

Fig. 86.

In summer time, when the mean vertical temperature gradient is AB, Fig. 86, the vertical temperature gradient of the greater mass of the descending anticyclonic air nearly coincides with the adiabatic line CD, but the air near the ground has strong diurnal range of temperature, as indicated by the downward divergence of the dotted lines DR and DN; rising to a high temperature by day, and cooling rapidly at night. The low nocturnal temperature generally produces fog in the lowlands; but the heat of the surface air in the day-time soon dissipates the fog and causes active local convection, generally producing fair-weather

cumulus clouds; the heat is sometimes excessive enough to give rise to local thunder storms, which will be considered in the next chapter. At night, the clouds all dissolve away, the air is habitually calm, and owing to the rapid cooling of the lower air by nocturnal radiation, there is a distinct inversion of temperature perceptible between valleys and hills; and the minimum on anticyclonic nights generally furnishes the lowest temperature of the month.

In winter time, anticyclones present very different conditions. The days then are short and the sun rises little above the horizon; insolation is brief and weak, and the temperature of the lower air in the clear space of anticyclones is dominated chiefly by radiation. The ground at such times, especially if snow covered, becomes excessively cold and the air near it is greatly cooled by radiation and conduction. The baric gradients being faint, the air lies nearly quiet and becomes colder and colder as long as the anticyclone endures. Thus the low temperatures imported by a cold wave are often followed by still lower minima locally produced in an anticyclonic calm. The cooling may become so excessive that, in spite of the small absolute humidity of the air, its dew-point is reached and then its lower layers are charged with a cold chilling ground fog. It is however only on land and in the lower air that these extreme features are developed. In the middle layers of the atmosphere, the air is neither cold nor foggy. Recalling the slow descent of the anticyclonic air and assuming AB, Fig. 87, as the mean vertical temperature gradient of our winter season, it follows that the vertical temperature gradient in winter anticyclones, above the reach of the stronger influence of the cold surface of the earth, will, as in summer anticyclones, approach close to the adiabatic line, CD, and thus produce an extraordinary contrast between the temperature and humidity of the middle and lower air, illustrated by the excessive inversions of temperature of such occasions. The air descending from altitudes above the limit of the figure has at heights above 2,500 feet a temperature gradient such as EH, which departs but little from the adiabatic gradient, CD; this indicates a temperature decidedly above the normal of the season. Mountain observations show that the increase of temperature may take place during the night, and without perceptible movement of the air. Nearer the ground, the tendency to

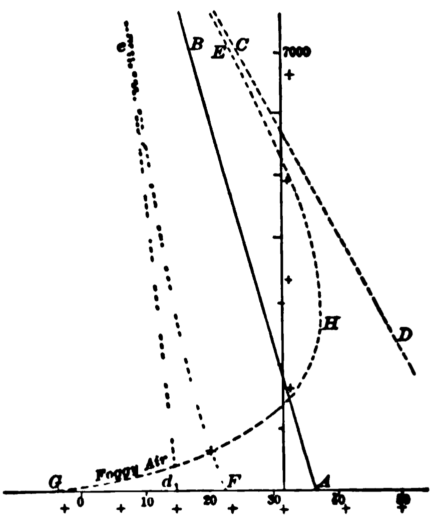

Fig. 87.

increase of temperature by compression during the slow and almost stagnating descent of the air, is overcome by radiation and conduction to the extremely cold surface of the ground, and the gradient becomes *HG*. The upper air being extremely dry, we may suppose that at a height of 7,000 feet, where the temperature is *E*, the dew-point is *e*. The rise of the dew-point in the descending air on account of compression alone would cause it to change during descent at the slow rate, *ed*; but if a slight addition of humidity is caused by the diffusion of vapor upward into the slowly settling air, the dew-point will rise at the faster rate, *eF*. The intersection of *EHG* and *eF* will indicate the altitude and temperature of the dew-point in the lower part of the anticyclonic mass; and below this the air will be foggy. The radiation which before took place from the ground is then continued, perhaps less effectively, from the upper part of the fog stratum, and its thickness increases. Thus while an observer on a mountain peak finds the air in winter anticyclones of relatively mild temperature and pleasant dryness under a clear sky, he may look down on a sea of fog submerging the lowlands, where all is cold, chilling, damp, and dark. Unlike the fair-weather valley fogs of summer nights, these winter fogs survive the day-time also, and may endure for a week in regions where the progression of the anticyclones is slow, as in central Europe.

It should be remembered that the strong contrasts of temperature here considered are not alone the result of cooling in the lower air, but of warming in the middle layers as well; and further, that the mountains, on which observations at higher levels are generally made, have nothing to do with producing the contrast, but serve only as convenient points for observation. Indeed, the mountain mass must serve to diminish somewhat the temperature of the air that settles upon it; observations from balloons would probably give stronger contrasts than those known from mountain stations; but balloon ascents are seldom made in the winter season.

Several examples of anticyclonic inversions in winter may be introduced. On December 27, 1884, when the pressure was high and the wind over New England was everywhere light, the early morning temperature on Mount Washington was $+16°$, while the records in the surrounding lowlands ranged from $-10°$ to $-24°$; thus showing that the reversed curvature of the line *EHG*, Fig. 87, is probably not exaggerated. Even Pike's Peak (14,134′) shows occasional higher winter temperatures then Denver (5,291′) (see Fig. 12, *c*, *d*), and these appear in all cases to be associated with anticyclones.

The most remarkable example of anticyclonic inversions of temperature yet studied occurred in Europe in December, 1879. The pressure of the month, reduced to sea level, averaged over 770 mm. (30.32 in.) over Austria, Germany, Switzerland and France, with lower pressures on all the surrounding regions; thus indicating persistent anticyclonic conditions for this period; and

the temperature in central Austria and Bavaria for the same time averaged less than — 10° centigrade (+ 14° F.). The lowlands were shrouded in fog for much of the month, while the mountain stations reported persistently clear weather and mild temperatures, over twenty Fahrenheit degrees warmer than the valleys 5,000 feet below them. The view from certain mountain summits was described as extending over a broad, smooth sea of cloud which concealed all the low country, while all the lofty mountains and here and there the higher hills rose above it like islands.

It is evident that the temperature on low ground during an anticyclone differs from that in the presence of a foehn on account of the difference in the velocity of descent of the air in the two cases. The foehn rushes down so rapidly that it loses by radiation and conduction but little of the heat that it gains by compression; the anticyclonic air settles down so slowly that it cannot preserve the heat that comes from its compression, and therefore suffers its temperature to be controlled by processes of cooling as it approaches the earth. It is also evident that all velocities of descent should be represented in the great variety of weather conditions between the extremes of these two classes of phenomena. If a foehn occurs with but moderate velocity, the peculiar features of its class will be faintly developed. If a local breeze occurs for some reason in an anticyclonic area, presumably accompanied by a more rapid descent than usual of the overlying air, cooling by radiation will have less time to counteract warming by compression, and a rise of temperature would be noted. Precisely this case has been detected in Austria, and it doubtless will yet be discovered in this country and in the interior of Canada. Like the bora that may be expected on the slopes of the plateaus of Utah and Arizona, the warmer breezes within the extremely cold areas of anticyclones should be critically looked for by observers favorably situated for their recognition.

250. Comparison of the foregoing examples. The well marked features of the several classes of cyclonic winds and of the anticyclonic calm may now be briefly reviewed. The sirocco is warm because it comes from a warm region; it is dry if derived from arid regions, or moist if flowing from warm seas into cyclonic centers. The cold wave is cold because it comes from a cold region; it is relatively dry because its temperature rises as it advances. The bora is cold in spite of its descent, because it was unduly cold before the descent began. The foehn is warm and dry on account of its supply from high levels at moderate temperatures and its rapid descent to lower levels. The inversion of temperature in winter anticyclones is due to the slow descent of their central air; thus allowing the production of relatively high temperature and low humidity at middle elevations, and of extremely low temperature and high humidity at low levels. Although these various classes may be con-

nected with one another by intermediate examples, they are all easily recognized when well developed, and hence serve as convenient types with which our many kinds of weather may be compared. They are not normal members of the general circulation, but are products of disturbances that interrupt the steadier flow of the great body of the atmosphere; they do not involve the higher levels of the air, but are for the most part developed in greatest distinctness close to the surface of the earth on which we live.

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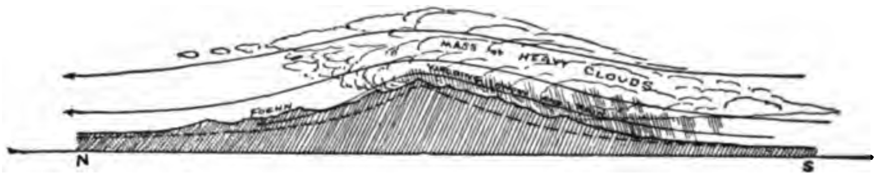

FIG. 84.

slopes disappears under its heated breath; hence the name *Schneefresser*, or snow-eater, sometimes locally given to it. Extensive fires among the wooden houses of the Swiss villages have happened at such times, the last one of the kind being that which destroyed Meiringen in the winter of 1891–92. The peculiar heat and dryness of the foehn are soon lost as it advances across the Piedmont plateau, cooling and absorbing vapor on its way.

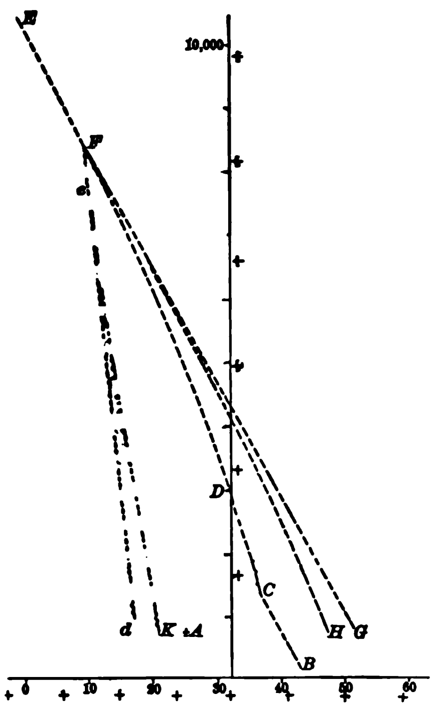

FIG. 85.

After the initiation of the foehn as thus described, it may be continued by a further supply of air coming from the plains of northern Italy, and then an additional cause for the heat and dryness of the wind is introduced. When the air is drawn away from the summit of the Alps, other air rises from the Italian lowlands, passes over the range, and descends on the leeward slopes, Fig. 84. Before ascent, the temperature of the air on the Italian plains may be represented by B, Fig. 85, while the temperature is A in the northern valleys. During the ascent of the Italian air, its temperature falls, its dew-point is reached, the whole mass becomes cloudy, and rain or snow falls from the clouds on the Italian slope. In the first part of the ascent, before the cloud level is reached, the temperature of the ascending air decreases at the rapid rate BC;[1] but when clouds are formed at the height C and above, further

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The increase of temperature produced by this peculiar reaction of latent heat will be stronger if the air is damp and relatively warm before beginning the ascent of the mountains, and under such conditions this second cause of the heat and dryness of the foehn may be as effective as the first; but it must be remembered that in the case of several foehn winds studied in Switzerland, the simple descent of the upper winter air is the first cause of the heat and dryness of the wind, and the liberation of latent heat is only a later and secondary cause; there sometimes being no rainfall on the southern slopes until a day or more after the fully-developed foehn is felt in the northern valleys.

The heat and dryness of the foehn are so unlike the cold and dampness of the wind on the mountain passes that it was only natural for earlier observers to ascribe the origin of the warm wind to some warm source; and it was consequently referred to the hot Sahara and regarded as an extension of the sirocco of southern Italy. This has been completely disproved. Espy and Dové were among the earlier meteorologists who suggested that the changes of temperature in vertical currents and the liberation of the latent heat from condensing vapor must be considered in its explanation; and this has been fully confirmed by modern students; especially by Hann of Vienna, to whom the suggestion of the initial cause of the heat and dryness of the foehn is due, and whose studies

have given careful demonstration of the more general suggestions of others. When thus explained, it is manifest that the foehn need not be limited to a south wind in the northern valleys of the Alps; it might occur under fitting conditions as a north wind in the southern valleys of those mountains, and such has proved to be the case; it might occur at the base of other mountain ranges, as has been abundantly shown. Wherever a lively cyclonic indraft draws the air away from the valleys at the foot of lofty mountains, leaving them to be filled by the descent of air from the altitude of the mountain crests and later by the passage of air over the range, a foehn-like wind may be expected.

These conditions are admirably developed along the eastern base of our Rocky Mountains in Montana, Wyoming, and Colorado, as well as in the Northwest Territories of Canada. It frequently happens in the winter season that as a cyclonic center moves eastward from British Columbia to Manitoba, while an anticyclone follows across Utah, an extended cyclonic circulation is developed over the mountain region. The winds that blow northward along the plains, as the cyclone advances, are soon supplanted by westerly winds; and as these descend from the mountains and flow out upon the plains, all the features of the Swiss foehn are developed. The warm and dry wind thus produced over a belt of country along the foot of the mountains is called the *chinook*. While the west wind may be damp and chilly under heavy clouds with a plentiful fall of rain or snow on the western side of the Front Range, the sky is fair or clear over the plains, and the clouds are left behind at the summit of the mountains; and although the velocity of the wind east of the mountains may be high, and its arrival may be at night, its temperature is mild or even warm, in marked contrast with the colder air that it has displaced and with the cold wave that usually follows a few days later. The warm chinook may arrive at night as well as by day (see Fig. 10, curve *g*); it quickly melts or dries up the snows of preceding storms, thus laying bare the northern plains and enabling cattle to survive the winter without protection. An isolated area of relatively high temperature may thus be produced by an extended chinook along the eastern base of the Rocky Mountains. Owen's Valley, below the Sierra Nevada in eastern California, should possess well-developed chinook winds. It lies to leeward of a lofty mountain range whose western slope is visited by heavy snow storms in the winter season; but as yet no accounts of such winds have been received from that nearly desert region.

One of the most remarkable examples of the foehn is found on the western coast of Greenland, where the cyclonic wind descending from the icy plateau of the interior becomes unduly warm and dry, raising the temperature at sea level even thirty or forty degrees above the winter mean. One of the best examples of its occurrence was during nine consecutive days in November and

December, 1875, when it was as warm in western Greenland as in northern Italy, and warmer than in Canada, Iceland, Great Britain, and over the intervening Atlantic; at one station of observation it was warmer in the darkness of the polar night than in France at noon-day.

The southern hemisphere has well-marked foehns in New Zealand, where they descend from the mountains and blow as "northwesters" across the Canterbury plains; and again at the eastern base of the Andes in the Argentine Republic, where their occurrence presumably on the equatorward side of passing cyclones is symmetrical with that of the chinook in our western states.

249. The anticyclonic calm. Although it is not at first apparent why calm air should be associated with a group of winds, still under a natural classification it seems best to place anticyclones, in which the winds are very gentle or at rest, in close association with the more boisterous cyclonic movements of the atmosphere just considered. Anticyclones possess a gentle descending movement of the air, and a quiet marginal outflow near the earth's surface, as has been explained in Section 238. They are in consequence prevailingly clear or fair and dry atmospheric regions; and in this respect they are strongly unlike the cloudy and wet air of cyclonic areas. But as their movements are of a definite kind, so are their various other features; and hence they deserve consideration as phenomena having persistent and recognizable characteristics. Being prevailingly clear, their surface air takes on the features of their season, and they must therefore be considered under different heads for winter and summer.

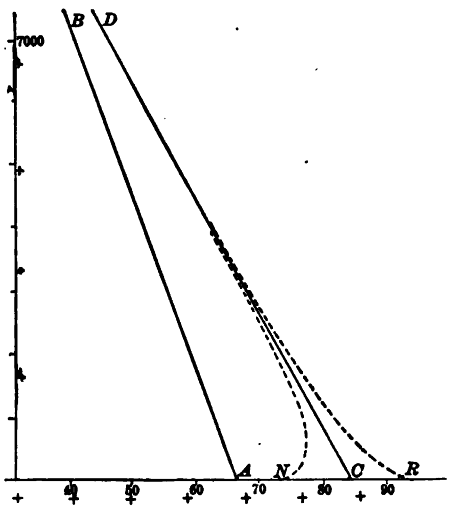

Fig. 86.

In summer time, when the mean vertical temperature gradient is *AB*, Fig. 86, the vertical temperature gradient of the greater mass of the descending anticyclonic air nearly coincides with the adiabatic line *CD*, but the air near the ground has strong diurnal range of temperature, as indicated by the downward divergence of the dotted lines *DR* and *DN*; rising to a high temperature by day, and cooling rapidly at night. The low nocturnal temperature generally produces fog in the lowlands; but the heat of the surface air in the day-time soon dissipates the fog and causes active local convection, generally producing fair-weather

cumulus clouds; the heat is sometimes excessive enough to give rise to local thunder storms, which will be considered in the next chapter. At night, the clouds all dissolve away, the air is habitually calm, and owing to the rapid cooling of the lower air by nocturnal radiation, there is a distinct inversion of temperature perceptible between valleys and hills; and the minimum on anticyclonic nights generally furnishes the lowest temperature of the month.

In winter time, anticyclones present very different conditions. The days then are short and the sun rises little above the horizon; insolation is brief and weak, and the temperature of the lower air in the clear space of anticyclones is dominated chiefly by radiation. The ground at such times, especially if snow covered, becomes excessively cold and the air near it is greatly cooled by radiation and conduction. The baric gradients being faint, the air lies nearly quiet and becomes colder and colder as long as the anticyclone endures. Thus the low temperatures imported by a cold wave are often followed by still lower minima locally produced in an anticyclonic calm. The cooling may become so excessive that, in spite of the small absolute humidity of the air, its dew-point is reached and then its lower layers are charged with a cold chilling ground fog. It is however only on land and in the lower air that these extreme features are developed. In the middle layers of the atmosphere, the air is neither cold nor foggy. Recalling the slow descent of the anticyclonic air and assuming AB, Fig. 87, as the mean vertical temperature gradient of our winter season, it follows that the vertical temperature gradient in winter anticyclones, above the reach of the stronger influence of the cold surface of the earth, will, as in summer anticyclones, approach close to the adiabatic line, CD, and thus produce an extraordinary contrast between the temperature and humidity of the middle and lower air, illustrated by the excessive inversions of temperature of such occasions. The air descending from altitudes above the limit of the figure has at heights above 2,500 feet a temperature gradient such as EH, which departs but little from the adiabatic gradient, CD; this indicates a temperature decidedly above the normal of the season. Mountain observations show that the increase of temperature may take place during the night, and without perceptible movement of the air. Nearer the ground, the tendency to

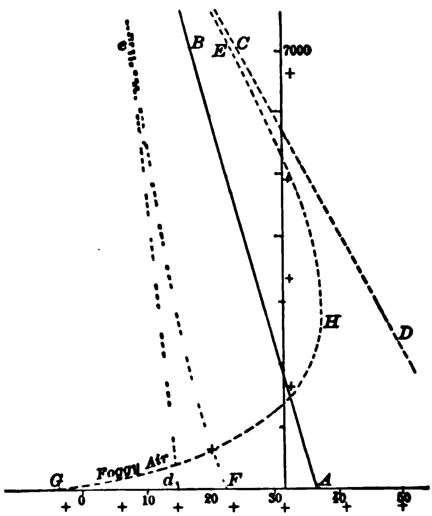

FIG. 87.

increase of temperature by compression during the slow and almost stagnating descent of the air, is overcome by radiation and conduction to the extremely cold surface of the ground, and the gradient becomes *HG*. The upper air being extremely dry, we may suppose that at a height of 7,000 feet, where the temperature is *E*, the dew-point is *e*. The rise of the dew-point in the descending air on account of compression alone would cause it to change during descent at the slow rate, *ed*; but if a slight addition of humidity is caused by the diffusion of vapor upward into the slowly settling air, the dew-point will rise at the faster rate, *eF*. The intersection of *EHG* and *eF* will indicate the altitude and temperature of the dew-point in the lower part of the anticyclonic mass; and below this the air will be foggy. The radiation which before took place from the ground is then continued, perhaps less effectively, from the upper part of the fog stratum, and its thickness increases. Thus while an observer on a mountain peak finds the air in winter anticyclones of relatively mild temperature and pleasant dryness under a clear sky, he may look down on a sea of fog submerging the lowlands, where all is cold, chilling, damp, and dark. Unlike the fair-weather valley fogs of summer nights, these winter fogs survive the day-time also, and may endure for a week in regions where the progression of the anticyclones is slow, as in central Europe.

It should be remembered that the strong contrasts of temperature here considered are not alone the result of cooling in the lower air, but of warming in the middle layers as well; and further, that the mountains, on which observations at higher levels are generally made, have nothing to do with producing the contrast, but serve only as convenient points for observation. Indeed, the mountain mass must serve to diminish somewhat the temperature of the air that settles upon it; observations from balloons would probably give stronger contrasts than those known from mountain stations; but balloon ascents are seldom made in the winter season.

Several examples of anticyclonic inversions in winter may be introduced. On December 27, 1884, when the pressure was high and the wind over New England was everywhere light, the early morning temperature on Mount Washington was +16°, while the records in the surrounding lowlands ranged from −10° to −24°; thus showing that the reversed curvature of the line *EHG*, Fig. 87, is probably not exaggerated. Even Pike's Peak (14,134′) shows occasional higher winter temperatures then Denver (5,291′) (see Fig. 12, *c*, *d*), and these appear in all cases to be associated with anticyclones.

The most remarkable example of anticyclonic inversions of temperature yet studied occurred in Europe in December, 1879. The pressure of the month, reduced to sea level, averaged over 770 mm. (30.32 in.) over Austria, Germany, Switzerland and France, with lower pressures on all the surrounding regions; thus indicating persistent anticyclonic conditions for this period; and

the temperature in central Austria and Bavaria for the same time averaged less than — 10° centigrade (+ 14° F.). The lowlands were shrouded in fog for much of the month, while the mountain stations reported persistently clear weather and mild temperatures, over twenty Fahrenheit degrees warmer than the valleys 5,000 feet below them. The view from certain mountain summits was described as extending over a broad, smooth sea of cloud which concealed all the low country, while all the lofty mountains and here and there the higher hills rose above it like islands.

It is evident that the temperature on low ground during an anticyclone differs from that in the presence of a foehn on account of the difference in the velocity of descent of the air in the two cases. The foehn rushes down so rapidly that it loses by radiation and conduction but little of the heat that it gains by compression; the anticyclonic air settles down so slowly that it cannot preserve the heat that comes from its compression, and therefore suffers its temperature to be controlled by processes of cooling as it approaches the earth. It is also evident that all velocities of descent should be represented in the great variety of weather conditions between the extremes of these two classes of phenomena. If a foehn occurs with but moderate velocity, the peculiar features of its class will be faintly developed. If a local breeze occurs for some reason in an anticyclonic area, presumably accompanied by a more rapid descent than usual of the overlying air, cooling by radiation will have less time to counteract warming by compression, and a rise of temperature would be noted. Precisely this case has been detected in Austria, and it doubtless will yet be discovered in this country and in the interior of Canada. Like the bora that may be expected on the slopes of the plateaus of Utah and Arizona, the warmer breezes within the extremely cold areas of anticyclones should be critically looked for by observers favorably situated for their recognition.

250. Comparison of the foregoing examples. The well marked features of the several classes of cyclonic winds and of the anticyclonic calm may now be briefly reviewed. The sirocco is warm because it comes from a warm region; it is dry if derived from arid regions, or moist if flowing from warm seas into cyclonic centers. The cold wave is cold because it comes from a cold region; it is relatively dry because its temperature rises as it advances. The bora is cold in spite of its descent, because it was unduly cold before the descent began. The foehn is warm and dry on account of its supply from high levels at moderate temperatures and its rapid descent to lower levels. The inversion of temperature in winter anticyclones is due to the slow descent of their central air; thus allowing the production of relatively high temperature and low humidity at middle elevations, and of extremely low temperature and high humidity at low levels. Although these various classes may be con-

nected with one another by intermediate examples, they are all easily recognized when well developed, and hence serve as convenient types with which our many kinds of weather may be compared. They are not normal members of the general circulation, but are products of disturbances that interrupt the steadier flow of the great body of the atmosphere; they do not involve the higher levels of the air, but are for the most part developed in greatest distinctness close to the surface of the earth on which we live.

CHAPTER XI.

LOCAL STORMS.

THUNDER STORMS.

251. Thunder storms and thunder squalls. Lightning is seen and thunder is heard in many rain storms that do not present the peculiarities of those to which the name, thunder storm, is best applied. Tropical hurricanes for example are accompanied by violent electrical manifestations near their centers, but they are not for this reason to be called thunder storms. Light falls of rain in the spring and summer are not infrequently in our country accompanied by moderate lightning and thunder; but these hardly deserve a stronger name than thunder showers. The typical and fully-developed thunder storm, with its peculiar outrushing squall or gust of wind, is a violent and relatively local disturbance, which certain well-marked features of cloud form distinguish clearly enough from other kinds of storms. Such storms occur chiefly in warm regions, in the warm season and in the afternoon or early evening.

252. The passage of a thunder storm. The coming of a well-formed summer thunder storm is heralded by a forerunning layer of cirro-stratus cloud, *c*, *c*, Fig. 88, commonly appearing in the west during the afternoon; fibrous or hazy at the forward edge, growing thicker and sometimes showing smaller or larger festoons (*f*, *f*), slowly descending and dissolving from its

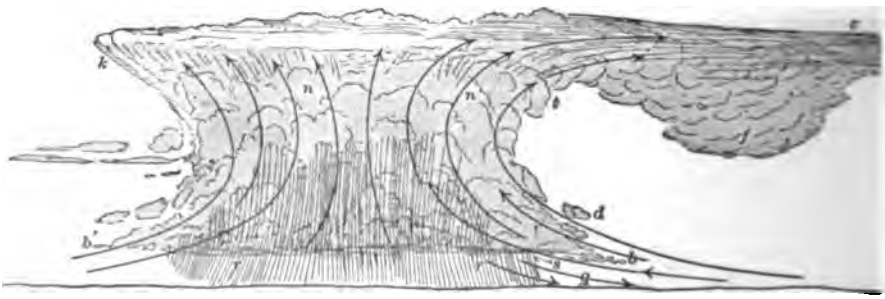

FIG. 88a.

under surface as the great rain-bearing cloud mass (*n*) approaches. A group of such festoons, observed near Philadelphia, July 16, 1887, is given in Fig. 89. The cirro-stratus cover may extend from ten to fifty miles in advance of the rain. The temperature preceding the storm is as a rule oppressively high, with light southerly winds, which have prevailed during an antecedent dry spell; but there is a slight cooling as the forerunning cloud cover spreads over

the sky and hides the sun. Perhaps an hour after the first sight of the cirro-stratus sheet, one may see heavy cumulo-nimbus clouds or "thunder heads" (*t*) of a dull leaden color and threatening appearance rise underneath it on the

FIG. 89.

western horizon. Distant thunder is heard as the thunder heads come nearer; and then their low level base (*b*) may be seen, below which the gray rain curtain (*r*) trails to the ground and conceals all objects behind it; the height of the clouds at their base being but an eighth or a tenth of their summit height. Smaller detached clouds (*d*) often form in front of the main mass, drifting into it and increasing in size as they coalesce with the storm cloud. The entrance of these detached clouds (*d*) and the flow of the lower winds (*b*)

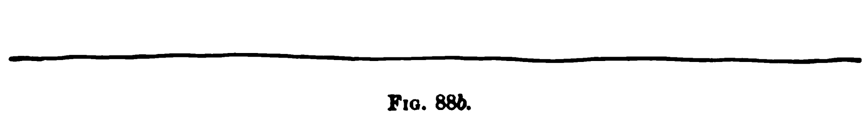

FIG. 88*b*.

into the great storm cloud does not necessarily imply a westward motion. They may move to the east; for the storm is advancing eastward and will overtake them if their movement is slower than its own. A ragged light gray "squall cloud" (*s*) rolls beneath the great dark cloud mass, a little behind its forward edge; and the whole structure advances broadside across country at

a rate of from twenty to fifty miles an hour. Below the clouds and in front of the rain is the short-lived outrushing wind squall (q), brushing up the dust that has been parched in the preceding drought; a cool blast in strong contrast with the relatively stagnant hot southerly air that preceded the storm. The temperature may fall ten or twenty degrees in twice as many minutes when the squall arrives, as appears in the tracing of a thermograph record, Fig. 11a, at Providence, R. I., for July 21, 1885. The barometer, that has been falling slowly up to this time, suddenly rises about five hundredths of an inch as the squall arrives; then the wind soon weakens, and the barometer after standing steady or falling slightly, gradually rises as the storm breaks away. The rain begins in large pelting drops shortly after the onset of the squall, and soon increases to a heavy downpour, often with hail; and at the same time the humidity rapidly increases almost or quite to saturation. The thunder, growing louder while the rain approached, now follows quickly after vivid flashes of lightning; and by this time the darkness of the shadow in front of the rain is already diminishing. The storm moves rapidly across country, and in half an hour, more or less, the rain slackens and the clouds break in the west. As they drift away, the pure blue sky is seen in the rear of the storm, and a little later the rainbow springs over the eastern horizon on the after side of the rain curtain, opposite the low sun near its setting in the west. The air is left cooler and cleaner than before; and the refreshed colors of the landscape, brightening towards sunset, form a most grateful contrast to the hot glare of noon and to the dark uproar of the storm. As the storm recedes and its thunder dies away, the rear of its festooned or bracketed overflow (k) may be seen far in the east, reaching somewhat backward from the top of the great cloud mass, delicately tinted by the rays of the setting sun. The clouds reach a mountainous height, and retain a pink glow after their base is lost in the dull blue sky underneath the twilight arch.

253. Observation of thunder storms. Observations at single stations serve to give the hours and seasons when thunder storms are most frequent; and it is thus found that a decided excess of storms occurs during the warmer spells of the warm season and in the afternoon or early evening hours; but this kind of observation does not suffice for the determination of their larger features. For this reason, the systematic study of thunder storms by numerous volunteer observers, all working on a uniform plan, has been attempted in the various countries of Europe and in different parts of the United States, with the most interesting results. The arrival of the squall wind, the beginning and ending of the rain, and the time of heaviest thunder, serve to mark the arrival and passage of the storm with much accuracy. The accounts of local storms, when reported for newspapers, should give at least some of these data, as well as a statement of the damage done by wind, rain, or lightning. When the records

from many stations are charted, and lines are drawn to indicate the place of the storm front at successive half hours or hours, a number of isolated storms of moderate size may be detected, all moving in a common direction; or a single large thunder storm may be found, advancing broadside or obliquely, with a belt of clouds fifty or a hundred or more miles in length and from ten to thirty miles in breadth, exclusive of the cirro-stratus cover which in large storms spreads out many miles in advance of the rain-cloud belt. The height of the cloud tops certainly reaches five miles in our stronger summer thunder storms, the upper clouds being then of ice crystals in spite of the high temperature at the bottom of the atmosphere. The following diagrams taken from various sources illustrate well-marked examples of thunder storms in this country and abroad.

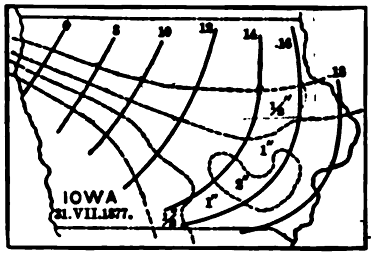

FIG. 90.

Fig. 90 is reproduced from the earliest map of a storm of this kind in the United States, made by Hinrichs no longer ago than July, 1877. It occurred in Iowa, where numerous local observations were used to define its advance. The storm front became wider as it moved forward; the heaviest rainfall lay along the axis of the storm, amounting to two inches in the southeastern part of the state; and the outrushing wind squall was of destructive strength at many places. A thunder squall, charted by Clayton, Fig. 91, began in northeastern Missouri about noon on July 5, 1884, and ran southeastward with convex front at a rate of a little over fifty miles an hour, until it faded away in northern Georgia about midnight. A small thunder squall of much intensity was observed to cross New England on July 21, 1885, Fig. 92, leaving the Hudson valley about ten o'clock in the morning, and reaching the ocean about an hour after noon. Its clouds were seen and its thunder was heard by observers to the north and south of its path where no rain fell. Beneath its dark shadow there was a rapid fall of temperature while its brief rain shower lasted, as illustrated in Fig. 11*a*; but southerly winds and high temperatures were resumed after its passage, and maintained until the arrival of a much larger storm in the late afternoon.

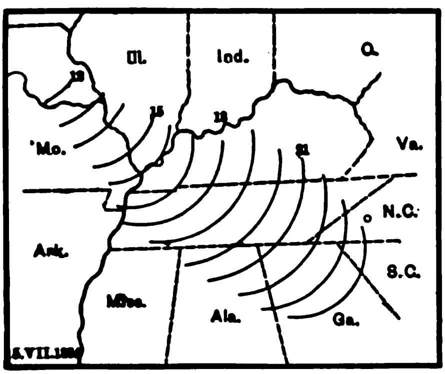

FIG. 91.

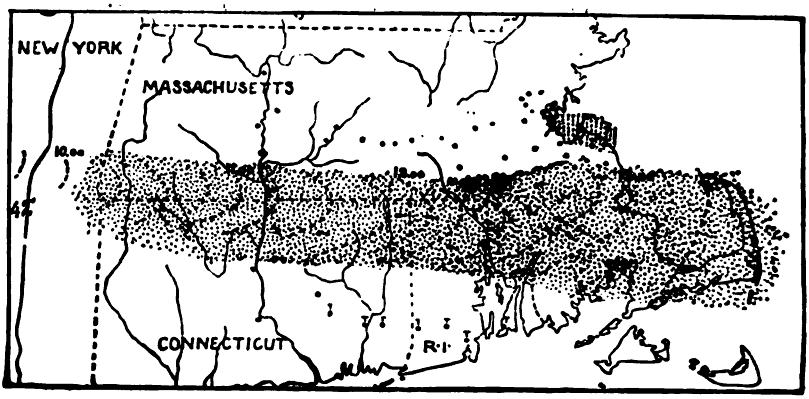

FIG. 92.

A remarkable thunder storm traversed northern and central Germany on August 9, 1881, the hourly positions of its somewhat discontinuous front being charted by Köppen in Fig. 93, from nine o'clock in the morning till nine in the evening. It gave few lightning strokes, but brought heavy rain, with hail at some places, and a violent outrushing squall, lasting five or ten minutes and doing much damage during its brief outburst. Light southerly winds with high temperature (85°) occupied the region in advance of the storm; cooler westerly winds (70°) followed it; the contrast of temperature being abrupt on either side of the rain front. As a rule, however, the thunder storms of Europe are less extended and less violent than those of the Mississippi valley.

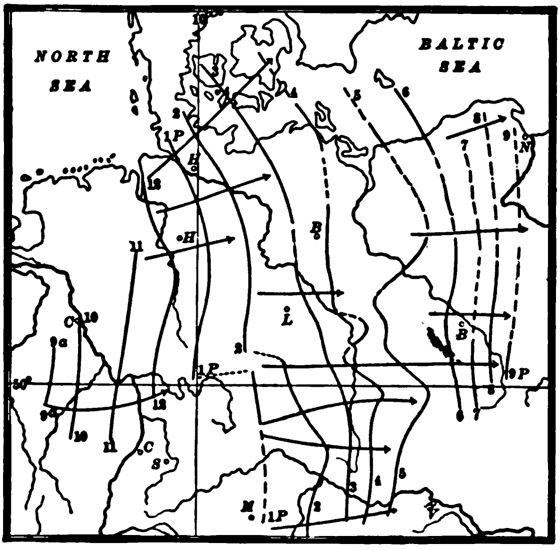

FIG. 93.

254. Convectional action in thunder storms. All the features of thunder storms point to their dependence on a convectional overturning of the atmosphere. They occur in warm regions and in the warm season, when

the vertical temperature gradient is stronger than in cold regions or in winter. They are most common and most violent in spells of warm summer weather, and at or shortly after the hour of the day when convectional movements are most active. They receive great assistance from the latent heat liberated from their abundant condensation of vapor at relatively high temperatures. Their early stages may be traced back to a beginning in ordinary cumulus clouds, in which the misty filaments are seen ascending and inflowing at the base, but rolling out and dissolving at the top on the side where the general motion of the air currents brushes them forward. Many such clouds may be watched during a summer morning from their first appearance, through their later growth to a large size, and then to their fading away; all these changes often requiring but a fraction of an hour, in which the cloud remains clearly in sight as it floats from west to east. The warmer the day, the larger the clouds and the longer their life.

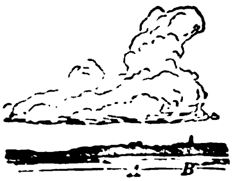

FIG. 94.

FIG. 95.

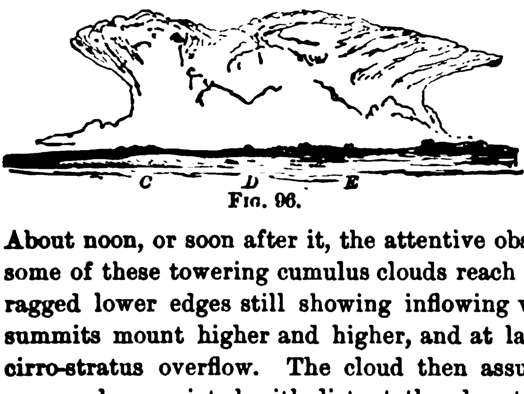

FIG. 96.

About noon, or soon after it, the attentive observer may sometimes notice that some of these towering cumulus clouds reach a greater size than the rest, their ragged lower edges still showing inflowing wisps, while the sharp-cut convex summits mount higher and higher, and at last manifest the initial stages of cirro-stratus overflow. The cloud then assumes the familiar anvil form, so commonly associated with distant thunder storms. Figs. 94 to 97 illustrate a series of such changes, by which an ordinary cumulus cloud is transformed into a thunder storm nimbus: they were sketched when looking northward from near New York city at 11.00, 11.15, 11.40, and 12.45 o'clock on July 2, 1887. The storm drifted to the northeast, and passed out of sight.


FIG. 97.

Even after the first cirrus overflow takes place, the cloud may fail to produce much rain, unless the process of growth is actively continued; but if the air be especially hot and sultry, a full development is likely to follow such a beginning. If an overflowing cloud of this kind comes in sight over

the western horizon shortly after noon, floating eastward in the upper winds, its further increase to mature size and strength may generally be seen as it passes the observer. The inflow at the forward lower edge is easily recognized by the movement of the cloud wisps; the ascent of the currents within the great cloud mass is clearly demonstrated by the upward expansion of the lofty thunder heads; the outflow at the summit is manifested in the cirro-stratus sheet, presumably beginning at an altitude where the ascending air is cooled to the temperature of the air around it. Sometimes one part of the cloud may reach a somewhat greater height than the rest, as if supplied by a rather warmer or moister indraft at the base. When the summit outflow is well established, it flows forward in the faster-moving upper currents; sometimes toppling over tumultuously, and dissolving away as it settles to lower levels; sometimes spreading evenly eastward in a broad sheet, more or less distinctly festooned on its under surface.

The first trails of rain fall from the base of the cloud and the first peals of thunder are heard from within it at about the time that the top of the cloud begins to spread out; presumably because the change from a nearly vertical ascent to a horizontal outflow fails to support the cloud particles; they fall through the cloud, increasing in size, and reach the ground as large drops of rain. The occurrence of hail is also indicative of active convectional motion, as will be more fully explained in Section 279. Sometimes many separate thunder clouds move eastward at a common speed; at other times, many clouds isolated at first seem to coalesce and form the long cloud belts of the large thunder storms described above. The constriction of the cloud mass between its broad base and its still broader overflow, producing the anvil form, has given rise to the idea that thunder storms consist of two cloud layers, between which the lightning flashes and the hailstones rise and fall; but this is now contradicted by many observations. The side view of distant storms shows their cloud mass to be continuous from base to summit; and observers on mountains or in balloons make no report of a clear space between the upper and lower levels of thunder storm clouds. The whole mass of the ascending current forms a single gigantic cloud. The same interpretation applies to cyclonic storms. Their cirrus overflow is often seen at a considerable height above their lower cumuliform clouds, with a space of clear air between; but this characterizes the marginal parts of the storm area, and nearer the center it is in the highest degree probable that the lower clouds are confluent upwards with the upper clouds, as illustrated in Fig. 60.

255. Geographical distribution of thunder storms. Thunder storms are common in the doldrums all around the equatorial regions. Here they generally occur in the afternoon or early evening, yielding heavy rain, by which the ocean surface is appreciably freshened for a time and its average salinity

is decreased as compared with that under the steady-going trade winds (see Fig. 105). Although active while they last, the thunder storms of the mid-ocean doldrums do not attain the terrific violence reached by thunder storms on or near the equatorial lands, such as are experienced in equatorial Africa and on the Atlantic waters to the west of the continent. As far as these are described, they seem to have, in a more intense degree, all the features of the stronger thunder squalls of our summer season. They are most violent in the late afternoon or evening. They possess massive clouds, drifting along in the higher winds; and hence advancing westward and passing from the land to the sea. They are surmounted by a cirro-stratus outflow, as with us. On their arrival, the wind suddenly shifts and blows out from beneath the clouds with great fury; and for this reason they are called African tornadoes.[1] Their rain is heavy and is accompanied by blinding flashes of lightning and a deafening roar of thunder. They last only a short time, and as they pass away, the atmosphere returns to its orderly diurnal changes, by which the torrid zone is so strongly characterized.

African tornadoes advance broadside over the ocean after the manner described above for our thunder storms; their long front, stretching across the sky, seems to be higher in the middle than at the ends; hence the name, "arched squall," often applied to storms of this class. In India, storms of the same character sweep down the plains of the Ganges from the west or northwest in the hot season, and are there called "nor'westers." Further inland, towards the desert region of the Indus, the amount of rain falling in such storms diminishes; but the clouds are formed, and the squall rushes out from beneath them, raising a dense cloud of dust from the parched ground; and these overturnings are consequently called "dust storms."

It is probable that the simoom of the Sahara and of Arabia, although without rain, clouds and thunder, is of similar convectional origin, as it is characterized by a rapidly-advancing wind, drifting the surface sands and raising great volumes of dust over the deserts. This, however, is a hot wind, there being no rain to cool it, and its temperature being greatly increased by the heat taken from the drifting sand as well as by strong sunshine. It has overwhelmed caravans, suffocating both men and beasts; but there is no reason to believe in its supposed poisonous qualities. The excessive heat, dryness and dustiness of the air are its dangerous features. It should, however, be added that certain hot winds in deserts, described as simooms, are not explained simply by referring them to convectional overturnings, like thunder storms except for their dryness.

In the Argentine Republic and Uruguay, thunder storms of great energy are observed in the summer season, under the name of "pamperos." Their heavy clouds are preceded by the long cirro-stratus cover; their nearer advance

[1] See note on this name, Section 266.

discloses the broadside approach of the lower horizontal cloud front, apparently arching from horizon to horizon; the squall brushes up a frightful cloud of dust from the dry pampas, and this is shortly followed by drenching rain with incessant lightning and thunder. These storms are greatly dreaded by ship masters in the estuary of the Rio de la Plata.

The cloud-bursts[1] of our arid western districts are only exaggerated thunder storms. They are local and short-lived, and seem to result from the sudden overturning of a large mass of unstable atmosphere. The clouds that accompany these storms have every feature indicative of a convectional origin, and, as with us, may be placed at the end of a well-continued series, beginning with ordinary cumulus clouds; passing then to moderate thunder showers, from which so little rain falls that it evaporates on its way down through the thirsty lower air, and hardly a drop reaches the parched ground; next to more active local thunder storms of the usual type; and all these culminating in the drenching fall of waters from the cloud-burst. A narrow strip of country is inundated by such storms for a short distance; temporary streams then rush down channels that are nearly dry at other times, gathering sand and dust, and delivering the discharge of the storm to the main valleys in dark, muddy torrents, many miles from the place of the rainfall.

258. Mountain thunder storms. The ascending valley breezes that run up the slopes of mountains by day frequently become energetic enough in the summer season to form clouds above the mountain summit, and afternoon thunder storms are often generated in this way. They drift away from the mountain over which their formation began, and the rain that falls from them trails down to the lower ground; but they seldom survive long, unless other conditions favor their growth. Storms of this kind are well known in the Alps and in the mountains of our western territory. It is not uncommon for an observer in the desert plains between the mountain ranges of Arizona to see several active thunder showers over the higher peaks; their rain may cause a rise in streams from the mountains and enable them to creep further down into the desert before being lost in the dry sands; but very little rain falls on the plains from such storms.

An interesting example of the combined action of the diurnal sea breeze and the valley breeze in forming a thunder cloud was recorded by a sea-captain as long ago as 1815 on the mountainous island of Hawaii in the North Pacific ocean. He wrote that soon after the sea breeze set in, about nine o'clock in the morning, a cloud began to form on the mountain slopes, surrounding the lofty volcanic summit in the center of the island in the form of a ring, like the wooden horizon that surrounds the artificial globe; and rain was soon

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afterwards seen to fall in torrents from the cloud. This was continued over noon; but towards evening, when the sea breeze died away, the rain ceased and the cloud soon disappeared; the air then remained clear until the next morning, when the same sequence of changes would begin again with wonderful regularity. The mountain stood in bold relief, and from where the ship lay, a little off shore, the summit could always be seen above the cloud even when it was densest and blackest, with lightning flashing from it, as happened every day; the rain never extended beyond the base of the mountain, and all around there was a cloudless sky.

257. Relation of thunder storms to cyclones. The preparation of synoptic weather maps, such as those of Figs. 64 and 67, by methods more fully explained in Chapter XIII, has shown that well developed summer thunder storms occur in our country and in Europe with greatest violence and frequency in the southern or southeastern part of cyclonic areas, more or less independent of the cyclonic clouds and rain, and from two hundred to five or six hundred miles distant from the center of low pressure. The storms are especially strong when the isobars in this district turn outward from the center in a pouch-like curve. When the cyclonic area, as defined by its isobars, takes the form of a trough or "V," elongated to the south-southwest, as in Fig. 64, thunder storms are often generated near or east of the axis of the trough, as if along the line of more rapidly falling temperature, where the southerly winds are replaced by currents from the west.

The general features of this important relation are illustrated in Fig. 98, by Hazen. The dotted line crossing the Great Lakes marks the track of a cyclonic center; its place at 7 A. M. of May 18, 19, and 20, 1884, being indicated by the figures of these dates. A series of curved lines is drawn to locate the successive positions of the belt along which the many thunder storms were observed on these days; those of May 18 being broken lines; those of May 19 being full lines. The position of the belt at certain hours is indicated by numbers at the end of the curved lines, counting from midnight to midnight. It thus appears that while

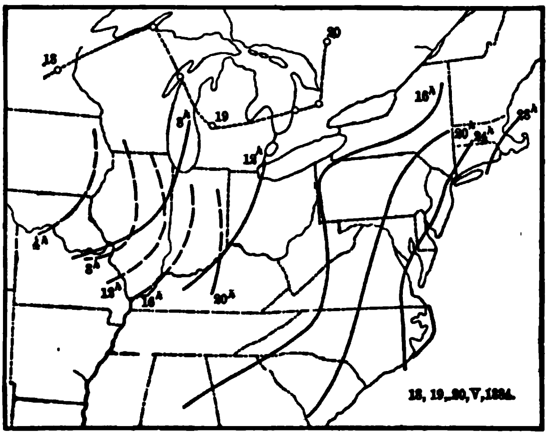

Fig. 98.

the cyclonic center was traversing northern Wisconsin, the belt of easy occurrence of thunder storms advanced from eastern Iowa and northeastern Missouri in the early morning, to eastern Indiana and central Kentucky in the evening; and that on the following day, while the cyclonic center crossed from lower Michigan into Canada, the belt first lay obliquely across Illinois, and then rapidly advanced eastward, increasing greatly in length as if influenced by the moister southerly winds nearer the coast, until at midnight it lay along the sea-board from Carolina to Connecticut. During all this time, a tolerably definite relation was maintained between the cyclonic center and the area of thunder storm occurrence. It is especially noteworthy that on May 19, the advance of the thunder storm belt was decidedly more rapid than the progression of the cyclonic center; as might be expected from the dependence of the belt on the winds which flow around the cyclone, and which therefore move eastward on the southern side of the cyclonic area faster than the progression of the center.

It is believed that this relation of thunder storms to cyclones gives further evidence of their dependence on atmospheric instability; but the instability is now seen to be determined not only by local heat due to sunshine on the day of occurrence, but also by the importation of air masses in the cyclonic circulation from different sources and with different temperatures. The following considerations will make this plain: Eastward of the cyclonic center, its inflowing winds partake of the nature of the sirocco; their vertical temperature gradient is represented by the line CDE, Fig. 99; they have a relatively high humidity when coming from the ocean, as has been explained in Section 245. Westward from the cyclonic center, the winds possess something of the characteristics of the cold or cool wave, with a vertical temperature gradient represented by the line FGH, and a relatively low humidity. It is not possible to state how greatly these vertical temperature gradients depart from AB, the mean vertical temperature gradient for the region and the season; but it is believed that their departures are of the kind here indicated. In the next place, numerous studies of cyclonic circulation have shown that the higher currents blow more to the right than the surface winds. It therefore follows that in a region between the district of the sirocco and the district of the cool wave, there may be a warm surface southerly wind overlain by a cool southwesterly wind; or a warm southwesterly wind overlain by a cool westerly

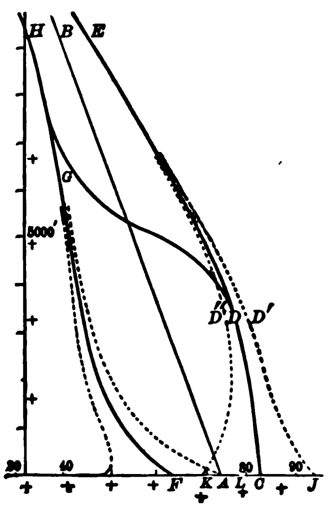

Fig. 99.

wind; and in this region, the vertical temperature gradient must have something of the peculiar form indicated in the curved line *CDGH*.

It is at present impossible to give any definite value to the gradient, *CDGH*, at various altitudes; but the considerations presented above make it extremely probable that in the region southeast of the cyclonic center, the southerly surface winds with high temperature and high humidity are overlain at a considerable altitude by westerly winds of lower temperature and lower humidity; and it is manifest that if any such arrangement exists, the opportunity for strong convectional overturning is thereby greatly increased over the opportunity arising only from the local heating of the lower air near the ground.

The occurrence of the greatest number of thunder storms at or shortly after the hottest hours of the day shows that the provocation to overturning that is caused by local and diurnal warming of the lower air is an effective assistance to the larger instability due to importation of unlike air-masses; the vertical temperature gradient then having values indicated by the line *JD'GH*. On the other hand, if a decided instability is caused by importation, thunder storms may continue to develop after nightfall, when the cooling of the lower air has changed the value of the temperature gradient to *KD''GH*. It is however manifest that theory has outstripped observation in this matter; and until additional facts are discovered by observation on mountains or in balloons, the discussion of the special value of vertical temperature gradients in cyclonic areas need not be pursued further.

There is, however, another way in which the cyclonic importation of air masses from diverse sources and with different temperatures and humidities has been thought to aid in the production of thunder storms, particularly in the development of those storms that advance broadside with a long front. This is by causing a *lateral* instability in trough-like cyclonic areas, across the belt where the cool and dry westerly winds advance close to the area occupied by the warm and moist southerly winds. An example of a cyclonic trough of this kind is shown in Fig. 64. It is probable that, in this case as in others, the higher members of the westerly currents overrun the southerly winds for a considerable distance eastward of the line that separates the two winds at the surface of the earth; for isolated local storms frequently spring up within the area of the southerly winds, with warm air extending many miles west of their rain. But along the line or belt where the cooler wind invades the area occupied by the warmer wind, it seems as if the latter were raised by the *under*-running of the former, and thus caused to roll over on itself; the ascending portion of the roll being recognized by the formation of great thunder storm clouds. Storms of this kind may perhaps be distinguished by the persistence of the change of temperature and winds that they introduce. Many of our larger storms may be referred to this cause; and the German storm, Fig. 93, is thought to be of the same kind.

It follows from these explanations not only that the cyclonic circulation may afford especially favorable opportunity for the development of convectional thunder storms in its southern or eastern quadrant (in our hemisphere), but also that so good an opportunity will not be found in any other part of the cyclonic area. Nowhere else can the arrangement of the cyclonic indrafts produce so rapid a vertical or lateral decrease of temperature as in the region to the southeast of the center of low pressure.

The strength of the contrasts of temperature in cyclonic winds will of course depend chiefly on the atmospheric contrasts of the regions whence a cyclone draws its supply of air. In the torrid zone, where the distribution of temperature is remarkably equable, no definite distribution of thunder storms within cyclonic areas has been detected. On the great southern oceans, where the temperatures are comparatively equable and relatively low in summer, violent thunder storms are not reported; although when the numerous cyclonic storms of the southern temperate zone cross New Zealand in summer, thunder storms occur about the time that the winds change from northerly to westerly; that is, in the same attitude with respect to the cyclonic center as with us. There is good reason to think that the same relation obtains in the Argentine Republic. The thunder storms of both Europe and the United States are dominated by cyclonic control; but of these two regions, the latter has the more numerous and violent storms, because of the stronger contrasts of its southerly and westerly winds. It should not therefore be thought that cyclonic action always develops thunder storms, but that the local storms spring up in the larger storms chiefly when the winds of the latter come from regions of strongly contrasted temperature and moisture.

It must furthermore be carefully noted that the cyclonic control of thunder storms is not by any means absolute. We have already seen that, besides the larger thunder storms which are formed along the axis of trough-like cyclonic areas, there are numerous smaller thunder storms distributed somewhat arbitrarily in the area of the warm and moist southerly winds. Some of these may be independent of the instability that is supposed to result from overflow by cooler westerly currents; they may depend essentially on the local superheating of the already warm and moist winds as they advance over our summer lands, the vertical temperature gradient resembling *JD'E*, Fig. 99. Again, it is well known that thunder storms of moderate extent but often of considerable activity spring up within the area of the westerly or northwesterly winds, southwest or west of the cyclonic center, where no instability due to importation of unlike air masses can be suspected. These storms may be best explained as the result of the rapid warming and moistening of the cool, dry westerly current as it flows under strong sunshine over a region watered by the preceding rain, gaining a temperature gradient represented by *AGH*, Fig. 99. Great cumulus clouds, often overflowing at the top, but

without developing fully into thunder storms, are characteristic of this region. A storm of this kind occurred in southeastern Massachusetts on July 17, 1889, while its parental cyclone was far down the St. Lawrence valley. It caused a heavy fall of hail, by which several valuable cranberry crops were laid waste. Finally, it is not uncommon to encounter thunder storms within anticyclonic areas in summer time; and here the instability on which the overturning depends must be referred entirely to the local warming and moistening of the lower air by intense insolation under the clear anticyclonic sky, the vertical temperature gradient taking a value indicated by *RD*, Fig. 86.

258. The progression of thunder storms. The general advance of thunder storms eastward in the temperate zone and westward in the torrid zone, has already been stated. It appears from this that, like smaller clouds on the one hand and like larger cyclonic storms on the other, thunder storms advance chiefly by drifting along in the general currents of the atmosphere in which they are formed. Their relation to cyclonic storms finds further illustration in this respect, for they move in directions about at right angles to the line running to the cyclonic center; if they are formed in the southeastern quadrant, they generally advance towards the northeast; if in the southwestern quadrant, they move toward the southeast. A few examples have been detected in Europe north of cyclonic centers, and moving toward the west.

The velocity of progression of thunder storms, commonly from twenty to fifty miles an hour, is somewhat greater than that of the cyclonic centers that they accompany. Such a result might have been expected because the local storms are borne as a rule in the winds which flow around the larger storm centers in the direction of their progression. This is illustrated in Fig. 98. The average progression of the cyclonic center was, in this case, twenty-one miles an hour; of the thunder storms, forty-one miles an hour.

The course followed by thunder storms within the cyclonic area gives strong confirmation to the suggestions of the preceding section regarding the overflow of westerly currents above the southerly winds. Even the isolated thunder storms that are formed southeast of the cyclonic center, far within the area of the southerly winds, move eastward, being borne along by a high-level current in that direction.

The eastward progression of thunder storms in temperate latitudes constantly carries them from the warm air of day-time into the cooler air of the night. They commonly gather strength during the afternoon and weaken in the evening, generally ending before midnight after having traversed a path of two, four or six hundred miles in length. Occasionally they endure longer, and it is possible that some which have been observed in the early morning hours have lasted over from the previous day; but there is no decision yet reached on this matter.

There is a very wide-spread belief that thunder storms follow valleys. Local reports mention over and over again the apparent deviation of thunder storms from the higher ground occupied by the observer, in order to follow a river course on the north or south. It is possible that an additional strength may be given to cloud growth when it is supplied by damp air from low ground, and thunder storms may in this way grow towards valleys and weaken over hills. It may be that the stronger centers of action in a long thunder storm front seem to pass more commonly to one side than directly over the observer, and that this is then interpreted as "following a valley"; but it may be safely asserted that thunder storms as a whole move over hills and valleys without significant deviation from their direct course. The example given in Fig. 92 of the thunder squall which crossed New England on July 21, 1885, affords a striking illustration of the disregard of high and low ground; it crossed the Hudson valley, the Berkshire plateau, the Connecticut valley, and the highland next to the east without any perceptible regard for their strong relief. However this may be in mountainous regions, it is clear that the surface of the eastern United States seldom has a strong enough relief to influence the course of thunder storms. They move in air currents so vast that they are indifferent to the moderate inequalities of the land.

The suggestion made above as to the aid given by the moister air of valleys in the growth of thunder storms may have an important application regarding the place of their more frequent beginning. It is suspected that the storms of New England frequently form in the Hudson valley about noon, and therefore arrive in New England in the late afternoon; Bavarian thunder storms have similarly been referred to the lowlands of the middle Rhine as a place of beginning; but further observations are needed on this point

259. The thunder squall. The violent outrushing squall of cool wind that commonly precedes our stronger thunder storms has been regarded by some observers as contradicting the belief in the convectional character of the storm as a whole; but when the subordinate dimensions and position of the squall with reference to the other parts of the storm are properly perceived, this contradiction entirely disappears. The squall does not reach to a great height above the surface of the earth; its upper limit must be below the dark lower front edge of the storm cloud — perhaps half a mile or less above the ground — for there the wisps of the cloud front demonstrate an inflow with respect to the storm in the most unmistakable manner. The squall does not extend far beyond the front of the main cloud mass; further in advance of the clouds, the warm surface wind often blows obliquely or directly towards the approaching storm. Fig. 100 represents the lower front of a thunder storm, with the squall below it; showing the space bounded by the letters *dbqs*, Fig. 88, on a larger scale. The dimensions of the squall are therefore relatively small, even

though its violence is excessive. It is a forward outflow from the bottom of the storm, but its occurrence by no means indicates that the whole mass of air in the storm is descending.

The cause of the squall has been sought in the downward brushing of the air by the falling rain; but the squall sometimes occurs under storm clouds from which no rain falls. It has been ascribed to the descent of air that has cooled under the shadow of the cloud; but it occurs at night as well as by day. It is best explained, following a suggestion by Ferrel, as the result of a downward reaction from the upward expansion of the great mass of air involved in the storm cloud. It is therefore analogous to the seaward reaction of the

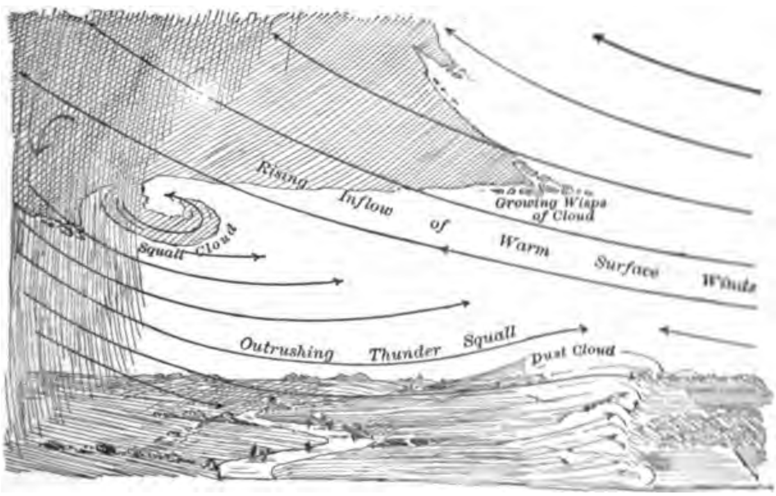

Fig. 100.

expanding air over a warm land area, by which the incoming of the morning sea breeze is delayed (Sect. 161). It may even be compared to the "kick" of a gun, and with more justice than at first appears. In the convectional overturning of cloudless air, the expansion of that which ascends is practically counter-balanced by the compression of that which descends; and hence there is no considerable change of volume as a result of the overturning. But in convectional action where clouds are formed, the volume of the air concerned in the overturning is actually greater after than before the change. The descending air is compressed at the usual rate; but the ascending cloudy air expands more than in the first case, because its cooling is retarded by the liberation of latent heat from its condensed vapor. To gain room for this increase of volume, a considerable mass of surrounding or overlying air must be pushed away by the ascending and expanding air; and in reacting on the ground it presses downward with more than its weight; thus causing at once

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An interesting example of the combined action of the diurnal sea breeze and the valley breeze in forming a thunder cloud was recorded by a sea-captain as long ago as 1815 on the mountainous island of Hawaii in the North Pacific ocean. He wrote that soon after the sea breeze set in, about nine o'clock in the morning, a cloud began to form on the mountain slopes, surrounding the lofty volcanic summit in the center of the island in the form of a ring, like the wooden horizon that surrounds the artificial globe; and rain was soon

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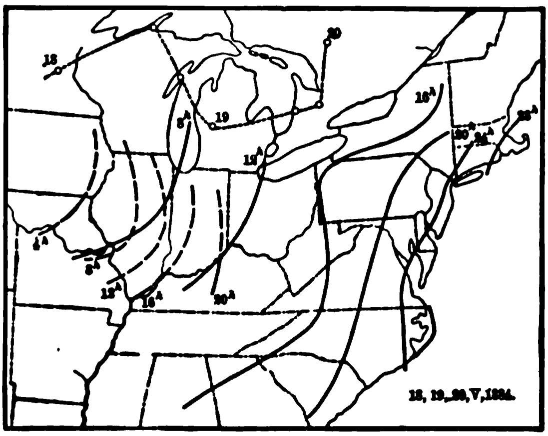

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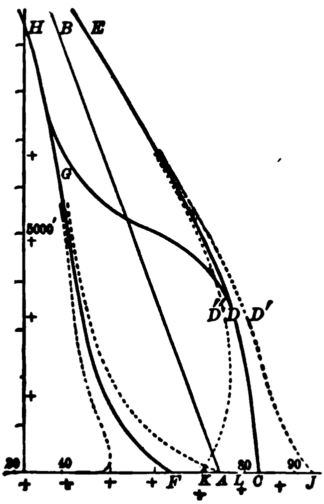

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without developing fully into thunder storms, are characteristic of this region. A storm of this kind occurred in southeastern Massachusetts on July 17, 1889, while its parental cyclone was far down the St. Lawrence valley. It caused a heavy fall of hail, by which several valuable cranberry crops were laid waste. Finally, it is not uncommon to encounter thunder storms within anticyclonic areas in summer time; and here the instability on which the overturning depends must be referred entirely to the local warming and moistening of the lower air by intense insolation under the clear anticyclonic sky, the vertical temperature gradient taking a value indicated by *RD*, Fig. 86.

258. The progression of thunder storms. The general advance of thunder storms eastward in the temperate zone and westward in the torrid zone, has already been stated. It appears from this that, like smaller clouds on the one hand and like larger cyclonic storms on the other, thunder storms advance chiefly by drifting along in the general currents of the atmosphere in which they are formed. Their relation to cyclonic storms finds further illustration in this respect, for they move in directions about at right angles to the line running to the cyclonic center; if they are formed in the southeastern quadrant, they generally advance towards the northeast; if in the southwestern quadrant, they move toward the southeast. A few examples have been detected in Europe north of cyclonic centers, and moving toward the west.

The velocity of progression of thunder storms, commonly from twenty to fifty miles an hour, is somewhat greater than that of the cyclonic centers that they accompany. Such a result might have been expected because the local storms are borne as a rule in the winds which flow around the larger storm centers in the direction of their progression. This is illustrated in Fig. 98. The average progression of the cyclonic center was, in this case, twenty-one miles an hour; of the thunder storms, forty-one miles an hour.

The course followed by thunder storms within the cyclonic area gives strong confirmation to the suggestions of the preceding section regarding the overflow of westerly currents above the southerly winds. Even the isolated thunder storms that are formed southeast of the cyclonic center, far within the area of the southerly winds, move eastward, being borne along by a high-level current in that direction.

The eastward progression of thunder storms in temperate latitudes constantly carries them from the warm air of day-time into the cooler air of the night. They commonly gather strength during the afternoon and weaken in the evening, generally ending before midnight after having traversed a path of two, four or six hundred miles in length. Occasionally they endure longer, and it is possible that some which have been observed in the early morning hours have lasted over from the previous day; but there is no decision yet reached on this matter.

There is a very wide-spread belief that thunder storms follow valleys. Local reports mention over and over again the apparent deviation of thunder storms from the higher ground occupied by the observer, in order to follow a river course on the north or south. It is possible that an additional strength may be given to cloud growth when it is supplied by damp air from low ground, and thunder storms may in this way grow towards valleys and weaken over hills. It may be that the stronger centers of action in a long thunder storm front seem to pass more commonly to one side than directly over the observer, and that this is then interpreted as "following a valley"; but it may be safely asserted that thunder storms as a whole move over hills and valleys without significant deviation from their direct course. The example given in Fig. 92 of the thunder squall which crossed New England on July 21, 1885, affords a striking illustration of the disregard of high and low ground; it crossed the Hudson valley, the Berkshire plateau, the Connecticut valley, and the highland next to the east without any perceptible regard for their strong relief. However this may be in mountainous regions, it is clear that the surface of the eastern United States seldom has a strong enough relief to influence the course of thunder storms. They move in air currents so vast that they are indifferent to the moderate inequalities of the land.

The suggestion made above as to the aid given by the moister air of valleys in the growth of thunder storms may have an important application regarding the place of their more frequent beginning. It is suspected that the storms of New England frequently form in the Hudson valley about noon, and therefore arrive in New England in the late afternoon; Bavarian thunder storms have similarly been referred to the lowlands of the middle Rhine as a place of beginning; but further observations are needed on this point

259. The thunder squall. The violent outrushing squall of cool wind that commonly precedes our stronger thunder storms has been regarded by some observers as contradicting the belief in the convectional character of the storm as a whole; but when the subordinate dimensions and position of the squall with reference to the other parts of the storm are properly perceived, this contradiction entirely disappears. The squall does not reach to a great height above the surface of the earth; its upper limit must be below the dark lower front edge of the storm cloud — perhaps half a mile or less above the ground — for there the wisps of the cloud front demonstrate an inflow with respect to the storm in the most unmistakable manner. The squall does not extend far beyond the front of the main cloud mass ; further in advance of the clouds, the warm surface wind often blows obliquely or directly towards the approaching storm. Fig. 100 represents the lower front of a thunder storm, with the squall below it; showing the space bounded by the letters *dbqs*, Fig. 88, on a larger scale. The dimensions of the squall are therefore relatively small, even

though its violence is excessive. It is a forward outflow from the bottom of the storm, but its occurrence by no means indicates that the whole mass of air in the storm is descending.

The cause of the squall has been sought in the downward brushing of the air by the falling rain; but the squall sometimes occurs under storm clouds from which no rain falls. It has been ascribed to the descent of air that has cooled under the shadow of the cloud; but it occurs at night as well as by day. It is best explained, following a suggestion by Ferrel, as the result of a downward reaction from the upward expansion of the great mass of air involved in the storm cloud. It is therefore analogous to the seaward reaction of the

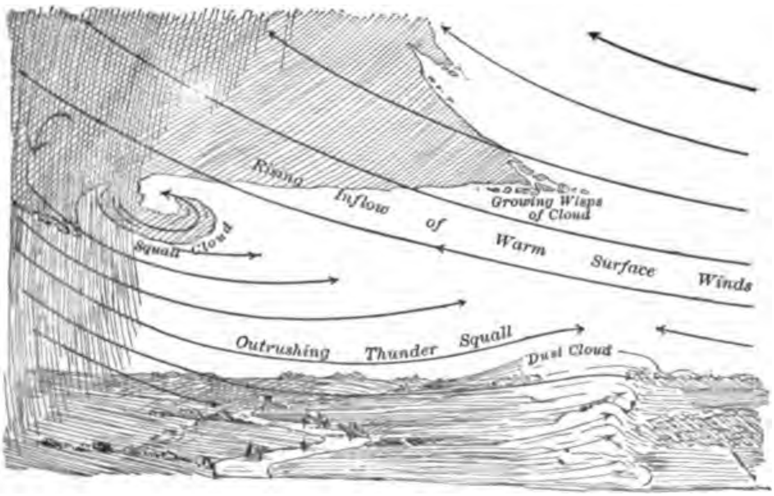

FIG. 100.

expanding air over a warm land area, by which the incoming of the morning sea breeze is delayed (Sect. 161). It may even be compared to the "kick" of a gun, and with more justice than at first appears. In the convectional overturning of cloudless air, the expansion of that which ascends is practically counter-balanced by the compression of that which descends; and hence there is no considerable change of volume as a result of the overturning. But in convectional action where clouds are formed, the volume of the air concerned in the overturning is actually greater after than before the change. The descending air is compressed at the usual rate; but the ascending cloudy air expands more than in the first case, because its cooling is retarded by the liberation of latent heat from its condensed vapor. To gain room for this increase of volume, a considerable mass of surrounding or overlying air must be pushed away by the ascending and expanding air; and in reacting on the ground it presses downward with more than its weight; thus causing at once

the slight rise in the barometer as the storm comes on, and the outrushing squall wind.

The development of the squall chiefly along the front of the storm, its less violence on either side, and its absence in the rear seem to be chiefly a result of the forward motion of the storm as a whole. The entire cloud mass floats forward at a rate of from twenty to forty miles an hour. Part of the air is pushed outward at the bottom. At the rear of the storm, the outward push is neutralized by the forward drift; moreover, the inflow that supplies the cloud is here relatively weak. In front of the storm, the two velocities of progression and expansion are combined, and the outflow thus becomes a destructive squall. This explanation is confirmed by the occasional occurrence of stationary thunder storms, in which the squall is felt with about equal violence on all sides of the base of the cloud.

A gray roll of cloud is generally observed at a little distance back of the dark lower cloud front; if carefully watched, it may be seen to turn slowly between the inflow above and the outflow below. Its presence seems to depend on an eddy caused by the squall, and it is therefore called the squall cloud, as in Fig. 100. A little way behind it, ragged cloud margins may be seen, from which the wisps settle down and dissolve away, as if brushed down by the squall wind.

260. Nocturnal thunder storms. A peculiar exception to the general rule of the occurrence of thunder storms in the hotter seasons and hours is found over the North Atlantic ocean in the middle and higher latitudes — and probably over other oceans in similar latitudes — and on the bordering coasts. In Iceland, for example, of twenty-three thunder storms recorded there in fourteen years, twenty-two were noted in the colder months and twenty were heard between sunset and sunrise. In Norway, most of the thunder storms occur on summer afternoons; but the smaller number observed in winter along the coast are more common at night than by day. The same rule holds good for western Scotland. Over the ocean, the excess at night seems to hold good; but the proportion during different seasons is not well made out. On our New England coast, winter thunder storms are relatively rare; but when they occur they are generally nocturnal.

The clear indication of convectional action in summer thunder storms on land leads to the belief that those of winter at sea should be convectional also; but in that case, their occurrence at night remains to be explained. It has been suggested that the reason for this may be found in the development of instability at such times by the excessive cooling of the upper layers of air by radiation from the lofty cyclonic cloud sheets in which winter marine thunder storms are developed; while the lower layers of air, on or near the ocean surface, are maintained at a comparatively high temperature. This

might be reasonably expected along the path of the Gulf Stream, or in warm winds derived from it. The resemblance of such a process to the development of the bora (Sect. 247) should be noted.

261. Atmospheric electricity. It is necessary to make a brief digression here in order to introduce some account of atmospheric electricity in general before speaking of its more intense manifestations in thunder storms. The air nearly always possesses a slight positive electric charge, compared with the earth. The source of this charge has been variously ascribed to the formation of vapor from water surfaces — but it is doubtful if the process of quiet evaporation suffices to account for it; to the friction of air on bodies that resist its motion, as dry air in dusty whirlwinds — but the quantity of electricity thus produced is small; to the friction of water particles and dry snow-flakes in the upper part of agitated clouds — but this process seems too exceptional to serve as the cause of so general a result. Further investigation is needed on this subject.

In studying the electrical condition of the atmosphere, it is found that the positive potential increases with altitude above the ground. It is subject to a slight, double diurnal variation, best observed in settled clear weather, having maxima in the morning and evening, and minima before sunrise and in the afternoon. It has also an irregular variation, accompanying changes in the weather; the positive charge being greatest under a clear sky, especially in the dry, cold air of winter anticyclones. In the presence of clouds, the charge varies greatly; sometimes becoming negative, or frequently changing its sign, especially in thunder storms, when its fluctuations are great and rapid. Its variations are therefore intimately connected with the quantity and condition of atmospheric vapor.

As a rule, the electricity of the atmosphere does not suffice to produce attractive or repulsive forces of sufficient amount to cause significant movements of the air. Exceptions to this statement may, perhaps, be found in thunder storms, but even there the general movement of the air seems to follow a convectional and not an electrical cause, as has already been described. Although many attempts have been made to explain local storms by electrical action, it has not yet been shown that the observed electrical forces nearly suffice to account for the results witnessed; nor has there been offered in such theories any reasonable and sufficient cause for the local development of electricity even in its observed insufficient amounts. It is manifest that a tenable electrical theory of storms must present a valid cause for the concentration of the electricity by which storms are to be produced; and that a theory is doubly at fault in calling upon an unexplained charge of electricity to produce results not accordant in quantity and quality with the observed effects of electric action. For example, in the working of a glass plate

electrical machine, a sufficient force must be applied to turn the plate before the sparks appear; so in the atmosphere, sufficient forces must be in operation to produce the intense charge that results in lightning flashes; but in both cases, the electric action is essentially the effect and not the cause of the other motions. The working forces of thunder storms are best accounted for by the convectional processes already described; and it is therefore more reasonable to place the lightning along with the clouds and rain, as secondary effects of the convectional storm, than to regard any one of them, itself not accounted for, yet serving as the cause of the others. While atmospheric electricity is an important branch of terrestrial physics, and while in thunder storms it attains an extraordinary display as an effect of the storm, it does not generally appear to be a factor of much importance in the processes of meteorology. In studying the general and local movements of the atmosphere, we are hardly more concerned with atmospheric electricity than with atmospheric composition.

262. Lightning. The identity of lightning with artificial electric sparks was suggested by Franklin shortly before the middle of the eighteenth century and demonstrated by experiment in France shortly after, as well as by Franklin himself in his famous experiment with a kite in Philadelphia in 1752.

When the growth of thunder storms is observed from their first appearance as cumulus clouds, it may be seen that lightning flashes first occur at about the time when the cirro-stratus sheet begins to spread out from the top of the cloud; and that from this time on the electric activity of the storm increases with its further cloud growth. The quantity of electricity in the cloud is continually increased by the inflow of moist air at its lower margin. It is believed that the electric potential is increased by the aggregation of many extremely small cloud particles into a smaller number of larger droplets, and finally into rain-drops; for the initial charge resides on the surface of each minutest particle, and with the successive aggregation of particles, the quantity of electricity increases faster than the surface area of the droplet. Thus with the growth of the cloud, there is both increased potential and increased quantity of electricity. It is probable that this process goes on in all cases of cloud formation; but that a potential high enough to cause lightning flashes is produced only when the cloud growth is rapidly and continually augmented by inflow of moist air at the base. Then the cloud droplets, suspended in the ascensional current of the cloud, gain a continual increase in both the quantity and potential of their electric charge, until a flash occurs. In our ordinary cyclonic storms, the cloud growth is gradual and the vertical component of movement is much slower than in thunder storms; hence there is less opportunity for the increase in the quantity of the charge by the process above suggested, and flashes are relatively rare. In tropical cyclones, where the

convectional process is much more active than in our latitudes, lightning is often frequent and vivid.

It is probable that when the different parts of a thunder cloud are thus variously charged, the lightning flashes depend for their opportunity in great measure on the movement of cloud masses and rainfalls within the storm. In this way, volumes of cloudy or rainy air of different charges may move about, rising or falling until they come within striking distance of one another, or of the earth. It is also suggested that the discharge of a flash may allow the union of many small droplets that were before held apart by electric repulsion, and thus locally promote the fall of rain. The heavier fall of rain that often reaches the ground shortly after a brilliant flash of lightning may perhaps be explained by either one of these theories; being regarded as the cause of the flash in the first, and the effect of the flash in the second theory. It is also possible that an intensely charged fall of rain may charge the particles near its path by induction, causing them to attract one another and thus promoting their coalescence and an increase in their potential.

Lightning flashes consist of several extremely brief sparks, each one almost instantaneous, separated from one another by small fractions of a second; hence the vibrating or flickering appearance of lightning often noticed. The composite nature of a flash is easily shown by looking at a thunder storm through a narrow slit in a rapidly-revolving disc; the slit being seen in several apparently stationary positions at each flash. The successive sparks are sometimes separately visible to the unaided eye. Each spark is believed to consist of many excessively rapid electric undulations, which cease when their energy is exhausted in overcoming the resistances on their path.

By far the greater number of flashes pass from one cloud to another. It is probable that they then begin and end in innumerable fine branches on countless cloud particles, and that the branches unite in the space between the clouds to form the single or composite trunk flash that we commonly observe. The brighter terminal branches are sometimes visible to the eye; but they are better perceived on a sensitive photographic plate, exposed at night in the direction of an active storm. Sometimes the flash passes from a cloud out to the open air, gradually dissipating itself. The length of flashes may reach several miles; but this has seldom been well determined. When a flash strikes the earth, it commonly selects some elevated point, such as a tree or church spire. The greater part of the discharge may then enter at a single point; but branches are often observed diverging towards various objects on the ground, and sometimes in great number, like the roots of a tree, or the fine endings of a nerve. A flash does not follow an angular zig-zag line, as it has been commonly represented in pictures; photographs show it to run in a sinuous path, somewhat like a river course. An apparently looped

or recrossing flash may be produced by the foreshortening of a twisted or helical path.

Lightning may be seen over great distances. Thunder storms in northern Italy have been witnessed in southern Germany, over the intervening Alps. The illumination of distant thunder clouds by inaudible lightning flashes is commonly called heat lightning; but this is not shown to differ from ordinary flashes, except in its distance from the observer. The occurrence of sheet lightning, reported by various observers, may be generally explained as the illumination of clouds by inaudible flashes; but it is possible that discharges in the thin upper air may be quiet and sheet-like, rather than noisy and disruptive.

Discharges of atmospheric electricity occasionally take the form of globe lightning, having the appearance of luminous balls, seeming to be a foot or so in diameter, moving at a moderate velocity and passing about among objects near the ground; remaining visible a number of seconds, and commonly disappearing with an explosion. No satisfactory explanation has been offered for this curious phenomenon: careful observation should be made of it. Weak brush-like discharges, known as St. Elmo's fire, are sometimes observed during stormy weather on trees and house tops, or on the yards and masts of vessels at sea. A similar appearance has often been noted on mountain summits within storm clouds; all pointed objects being surmounted by a bluish flame-like light, from which a buzzing or crackling sound is emitted. The fingers of an observer may thus discharge an electric stream into the air.

263. Thunder. The brilliancy of lightning is due to the excessive vibration of the luminiferous ether caused by the flash; the deafening sound of the thunder results from violent vibrations excited at the same time in the air. The sudden heating and electric disturbance along the path of the flash have much the same effect in producing sound as the firing of an explosive substance. When a flash occurs near the observer, the sharp crackling reports first heard come from the smaller branches that are nearer than the trunk; the heavy crash immediately following comes from the nearer part of the trunk flash; and the rolling thunder that then succeeds comes from the more distant part of the trunk, as well as from reverberation among the clouds. The rolling is greatly intensified among lofty mountains.

As sound travels through the air with a velocity of about 1,100 feet a second, the distance of a flash in miles is roughly equal to one-fifth of the number of seconds — or pulse beats — counted between the flash and its thunder. It is seldom that a longer interval than seventy or eighty seconds elapses between a flash and its sound. The velocity of progression of a thunder storm may be estimated by recording the time at which successive flashes appear, and determining the distance of each one. In doing this, it

will be noticed that the nearest flashes generally occur just after the beginning of the heavy rain; that is, near the front of the storm, where the inflow of moist air and the cloudy condensation of its vapor are most active.

284. Lightning strokes and lightning rods.[1] When lightning strikes the earth, it sometimes fuses the sand along its path, forming vitrified tubes or fulgurites, having a depth of several feet below the surface. A flash may pass along beneath the surface at a slight depth, turning up a furrow of earth, probably by the sudden vaporizing of the moisture that it encounters. In striking trees, the bark may be split off, or the trunk shattered; but if the bark is smooth and well wet by rain, little injury may be done. When an unprotected house is struck, its walls are more or less fractured, and if built of wood, it may be set on fire. There is no truth in the saying that lightning never strikes twice in the same place.

It is reported that in the six years, 1885–1890, there were 2,223 buildings set on fire by lightning in this country, or 1.3 per cent of the total number of fires; the loss thus occasioned being $3,386,826, or 1.25 per cent of the total fire loss. It is further reported that 205 persons were killed by lightning in this country in 1891, and 292 in 1892. Over 95 per cent of these casualties occur between April and September, or within half of the year. When a person is apparently killed by lightning, it may be that no permanent injury is effected, but that respiration is stopped by a temporary paralysis. Efforts to restore respiration should be continued for an hour.

All these disruptive and harmful effects are caused by the expenditure of the energy of the flash on poor or insufficient conductors that happen to lie in its path. It is therefore desirable to protect buildings by providing them with conductors or lightning rods, in which the undulating sparks of the flash may pass easily, leaving them to exhaust themselves elsewhere. Lightning conductors are best made in the form of continuous copper rods or tapes terminating upwards in one or several sharp points, ten or fifteen feet above the building; and extending downward beneath the building deep enough to be in permanently wet ground. The last point is extremely important; many rods fail to protect buildings by reason of the neglect of this essential in their construction.

It is found, however, that buildings are sometimes injured even when provided with good rods. The cause of this difficulty seems to lie in the variation of the intensity of flashes. Moderate discharges can be safely disposed of by ordinary rods; but excessive discharges overwhelm the conductors or their connection with the ground; and then destructive branching may occur through the building. A number of tall trees near a house probably afford better protection than most lightning rods.

[1] See "Lightning Conductors and Lightning Guards," by O. J. Lodge.

265. The aurora borealis, often called northern lights, is an illumination of the atmosphere in arches, streamers, patches or sheets of whitish, yellow, green or red light, caused by diffuse electrical discharges chiefly in the thin upper air. It is occasionally bright enough to be seen in the day-time. Although irrelevant to the chief subject of this chapter, some account of it is conveniently introduced in connection with atmospheric electricity, of which it is a peculiar manifestation.

The aurora is most common and brilliant in relatively high latitudes, along a belt that follows near the shores of the Arctic ocean from the North Cape of Europe eastward to Point Barrow in northwestern America; thence somewhat southward, so as to pass through Hudson Bay in latitude 60°, and a little south of Greenland; and obliquely northward again between Iceland and the Faroe. On either side of this belt, the aurora is less common; and in the torrid zone, it is rarely observed. The aurora australis is seen in high southern latitudes, but its distribution has been little studied for want of observations.

As commonly seen in our latitudes, the aurora begins with the formation of an arch, more or less complete, with its apex in the magnetic meridian; the lower side of the arch being better defined than the upper, and the sky beneath seeming darkened by contrast; but stars are visible there as well as through the aurora itself. The angular altitude at which the arch forms is greater in higher latitudes, until in the belt of greatest frequency the arch crosses the zenith, stretching at right angles to the magnetic meridian. Further towards the pole, it is seen to the south of the observer. The arch is sometimes evenly illuminated; sometimes convoluted like a folded curtain; but it is more commonly banded with rays nearly at right angles to its curve. In the greater displays, the rays are prolonged upwards into streamers, which seem to converge in a corona high in the sky, nearly in the direction indicated by the south end of a magnetic dipping needle. The light of the streamers often flashes rapidly, whence the name, "merry dancers," sometimes given to them. After its formation, the arch frequently moves slowly away from the direction of the belt of greatest frequency; that is, towards the equator in our latitudes, and towards the pole in the Arctic regions. As it moves, it has been noted that the apparent breadth of the arch diminishes on approaching the coronal point; that the streamers make a more and more acute angle with the arch until they coalesce with it when it passes through the coronal point; that the brightness of the arch is greatest when it is narrowest; and that on passing the coronal point, these changes proceed in the reverse order. It is concluded from these facts that the arch is like a sheet, hanging nearly vertical; and that the rays or streamers are nearly parallel, their apparent divergence from the coronal point being an effect of perspective, like that by which the beams of the sun shining between clouds, or the paths of a group of shooting stars are given an appearance of divergence.

It is believed that the auroral arch is a more or less extended arc of a circle whose plane is at right angles to, and whose center lies in the magnetic axis of the earth. Accepting this conclusion, the height of the arch has been found to vary between 33 and 281 miles, averaging 130 miles above the earth's surface. Its streamers seem to extend to even greater heights. On the other hand, faint rays have been reported as being visible between an observer and a neighboring mountain or a low cloud. Systematic observations by numerous observers are needed on this point.

Besides the geometrical forms of arches and streamers, the aurora has many irregular appearances, of which no account can be given here. Its duration varies greatly, from a faint light for a few minutes, to brilliant displays lasting many hours, or perhaps enduring over several days. These greater displays have been witnessed over large parts of the earth; while the ordinary lights are relatively local. The relation of auroral displays to atmospheric conditions, such as control weather changes, is not well made out, although they are commonly associated with fine clear skies. Certain Arctic observations, during long polar nights, indicate that the aurora is more frequent in the nocturnal than in the diurnal hours. It exhibits a double annual period, being more common in March and October, and less common in January and June. It also has a period of greater frequency about every eleven years; numerous aurorae corresponding with numerous sun-spots, and with the stronger disturbances of the magnetic needle; thus indicating some association of terrestrial magnetism and auroral displays with solar action. There is also a longer variation in the frequency of aurorae; they were relatively rare from 1795 to 1823, and relatively frequent about 1780–90, 1850 and 1870. While these various relations, both of place and time, leave no doubt that the aurora is an electric discharge chiefly in the upper air, and dependent in some way on the magnetic conditions of the earth and sun, the full nature of its controls and processes are by no means understood.

Tornadoes and Waterspouts.

266. Tornadoes are local whirlwinds of great energy, generally formed within thunder storms. Their most invariable feature is a funnel-shaped cloud that hangs from the bottom of the greater thunder cloud mass above. The funnel is created around the axis of a violent ascending vortex of whirling winds; its diameter may reach a few hundred feet, being much greater above than below; while the destructive winds around it cover a somewhat greater space. The whirling funnel advances generally eastward or northeastward, at a rate of twenty, thirty or forty miles an hour, with a deafening roaring noise, destroying everything that its winds fall on in their rapid passage. The activity of a single tornado may continue for half an hour or an hour,

while it lays waste a path five to twenty or more miles long, and commonly less than a quarter of a mile wide.[1] Waterspouts at sea are of essentially the same nature as tornadoes on land; but their association with thunder storms is less marked.

Although careful observations of passing tornadoes are seldom made, on account of the dread they naturally inspire, there are many accounts of them from observers who were at a safe distance on one side of their track; and from these we have repeated accounts of the whirling, writhing funnel cloud, which constitutes the visible part of the ascending vortex. An old account of a "spout" in England in 1587 is as follows: — "The wind thus blowing soon created a great vortex, giration and whirl among the clouds, the center of which ever now and then dropt down in the shape of a thick long black pipe, commonly called a spout; in which I could plainly and most distinctly behold a motion, like that of a screw, continually drawing upwards and screwing up (as it were) whatever it touched." A tornado in central Massachusetts in 1760 was thus described: — "At Leicester, several people of credit say that about five o'clock the sky looked strangely; that clouds from the southwest and northwest seemed to rush together very swiftly, and immediately upon their meeting, commenced a circular motion; presently after which a terrible noise was heard. The whirlwind cut its path through the trees, and after having passed over some clear land, it came to the dwelling of one David Lynde, the only one which stood in its way; upon this it fell with the utmost fury and in a moment effected its complete destruction." The following extract describes a tornado that was carefully observed in northern Alabama in 1867. When first seen, the funnel was about five miles away, and judging from its angular altitude above the horizon, its height was estimated at 4,500 feet; coming nearer, it passed about nine hundred feet south of the observer, when the gyratory motion of the cloud was distinctly visible. Many small objects gathered from the ground were perceived flying around the summit of the column like a great flock of birds; and a pine tree, afterwards found to be

[1] The name, tornado, was originally applied some two centuries or more ago, to the violent thunder squalls that frequent the western equatorial coast of Africa. It is of Spanish origin, although not a Spanish word, and refers to the rapid shifting of the wind on the outburst of the thunder squall. In this country, however, the word has been applied for about a century to the smaller whirling storms with pendant funnels, within larger storms. To add to the confusion of terms, our tornadoes are often called cyclones. Newspaper reports of destructive storms seldom call them by the proper name, and in many instances fail to give descriptions by which the true character of the storm can be recognized. In this book the name, cyclone, will be used only for large areas of low pressure, so well defined on the weather maps, with broad sheets of clouds, rain or snow of greater or less amount, and inflowing spiral winds, seldom of destructive strength on land, and occurring with us even more frequently and with greater strength in winter than in summer; while tornado, and its marine synonym, waterspout, will be used as described in this chapter.

sixteen inches in diameter and sixty feet long, was seen to float out from the black vortex, and sail around to all appearances as light as a feather.

The winds in the tornado vortex attain an incredible violence. Houses are torn to pieces, and their fragments are scattered for hundreds of feet along the track of the storm. Trees are torn up by the roots or broken from their stumps and stripped of their smaller branches. Men are carried violently through the air, falling at last with such force as to inflict fatal injuries or cause instant death. It is noteworthy however that the number of deaths reported in this country from violent winds in recent years is less than the number of deaths caused by lightning. Cattle have been impaled by flying boards. Heavy objects, such as plows, logs or chains, are carried many feet. Shingles, clothing and papers have been found a mile or more from where the wind caught them up. Chickens are stripped of their feathers. Nails are driven into boards. The variety of effects is endless; the scene of destruction produced by the passage of a tornado through a village is terrible in the extreme. The village of Grinnell, Iowa, was thus laid waste on June 17, 1882. Rochester, Minn., was devastated on Aug. 21, 1883. Many other examples of tornadoes, famous for their violence, could be added.

It is manifest from these accounts that whirling tornadoes and outrushing thunder squalls should not be confounded; although it is probable that many of the latter have been described under the former name. They are alike only in their associations with thunder storms, their suddenness and their brief duration. Tornadoes greatly exceed squalls in violence, while squalls greatly exceed tornadoes in the breadth of country over which they are felt as they advance. When both of these subordinate but violent winds occur in a single thunder storm, the squall would be the forerunner of the tornado; but the tornado would be felt on only a small part of the district over which the squall had swept.

267. Regions and seasons of occurrence. Tornadoes are more frequent in the Mississippi valley and in certain of the southern states than in other parts of this country. They have however been reported in all the states east of the great plains, and they are known in less frequent occurrence in Europe and other parts of the world. Being distinctly associated with thunder storms, tornadoes are found to occur in greatest number in the warmer months; but in the southern states they are sometimes reported during the warmer spells of the colder months. They are more frequent in the warmer afternoon hours and very few are recorded as occurring in the early morning.

Like thunder storms, tornadoes frequent the later part of warm spells; that is, the western part of areas of warm southerly, sirocco-like winds, or the southerly or southwesterly quadrant of cyclonic storms. It is by means of this relation of tornadoes to larger atmospheric disturbances that they may

some day be predicted in a general way, for nearly all the tornadoes that have been recorded since the preparation of our daily weather-maps have been found to lie south or southeast of a cyclonic center, at a distance varying from two to eight hundred miles from it. The prognostics of tornadoes are therefore sultry, moist, southerly winds bearing heavy clouds, often portentous in form, agitation and coloring.

The most remarkable example of the relation of tornadoes to cyclones yet noted was on February 19, 1884, when some forty tornadoes occurred in the southern states between morning and midnight. The morning weather-map of that day (Fig. 64) showed a trough-like cyclonic storm of considerable intensity central in Illinois. Its moist sirocco indraft was drawn from the Gulf over the southern states with a temperature of 50° or 80°; while the western states were occupied by west and northwest winds with a temperature near freezing, and in the far northwest even 10° or 20° below zero. All of the tornadoes reported on this disastrous day occurred near the western boundary of the area occupied by the sirocco. During the morning, they were reported in Mississippi and western Alabama. In the afternoon, the cyclonic center had moved over southern Michigan; and at this time the tornadoes were noted in eastern Alabama and Georgia. By evening, the cyclonic center had advanced to Lake Huron; tornadoes were then formed only in eastern Georgia and the Carolinas. Thus for the entire day, the district in which the violent whirlwinds were generated stood in a definite relation to the center of the larger cyclonic storm and its system of inflowing winds. The thunder storms of this day and the places of its heavier rains have not been specially studied; but it may be expected that their area marched eastward in the same manner as the tornado area.

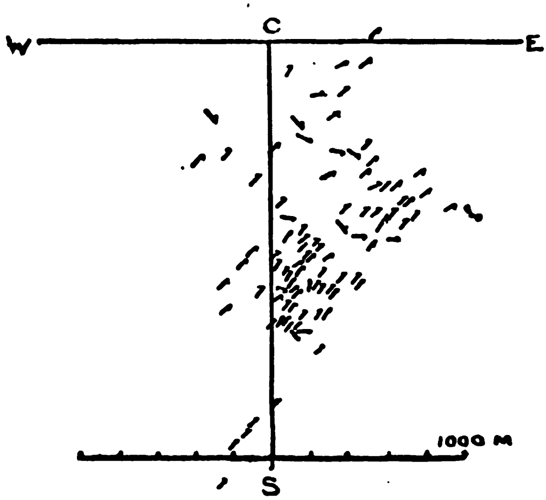

Fig. 101.

Many similar examples might be given, although none have the number of tornadoes reported on the disastrous day described above. In 1893, March 25, April 7 and 13 possessed warm sirocco winds in which numerous and destructive local storms were developed in the Mississippi valley.

Fig. 101 represents the positions of 135 tornadoes, recorded during 1884, with respect to the center of the cyclones in whose winds the smaller whirls

were formed. The direction of movement was not reported for those indicated by dots. It is apparent that nearly all of the tornadoes are limited to a definite space, south-southeast of the cyclonic center.

268. Convectional origin of tornadoes. In view of the prevailing association of tornadoes with the great cumulus mass of thunder storms, and of the

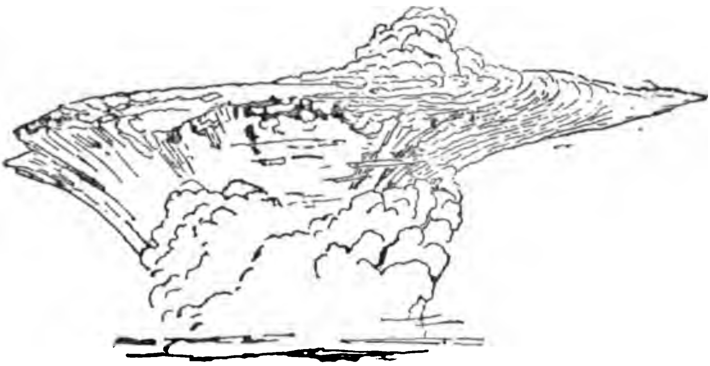

FIG. 102.

definite relation of tornadoes to cyclones, we cannot hesitate to refer them to some special form of convectional action, determined not alone by immediate and local sunshine at the warmer hours of the day, but more largely by the importation of air masses of different temperatures and humidities from diverse regions. Thunder storms have already been referred for the most part to the same opportunity, and it remains to be seen what difference of conditions shall determine the occurrence of thunder storms alone and of thunder storms with tornado funnels beneath them. The best suggestion yet offered for the development of tornadoes in thunder storms is based on the inferred occurrence of exceptionally strong updrafts here and there in the thunder clouds. A side view of a distant thunder storm often indicates the existence of locally strong updrafts by the rise of certain of the thunder heads to a greater height than the rest; as appears in the accompanying sketch, Fig. 102, drawn on August 26, 1886, looking northward from Philadelphia at a great thunder cloud whose distant base was lost in the hazy lower air. These updrafts are thought to depend in turn on the local occurrence of an unduly warm and moist mass of air, which may consequently ascend to great heights in the atmosphere. This supposition is further indicated by the prevailing association of tornadoes with thunder storms of great activity, in which the rain is heavier than usual, and from which hail not infrequently falls. The assistance derived from the liberation of latent heat from condensing vapor is of essential importance in this process; and it

must be remembered that the occurrence of condensation at high temperatures is particularly effective in retarding the cooling of ascending currents, and hence in aiding their ascent; because the decrease of vapor capacity is then so rapid.

It should be carefully borne in mind that the convection here considered does not depend on an atmospheric instability that is determined simply by the immediate and local warming of the lower layers of air by sunshine, even though this process may occasionally produce dust whirlwinds of somewhat destructive strength: nor does the instability depend on the long quiet brooding of the air day after day under strong sunshine, such as that which gives rise to the tropical cyclones in the doldrums. The instability that produces thunder storms and tornadoes is believed to depend on the importation of unlike masses of air from different sources into close neighborhood and into such relative positions that convectional overturning is a necessary result. Some meteorologists have questioned the sufficience of convection as a cause for the excessive violence of tornadoes; and have therefore appealed to the action of electricity or of some even more mysterious agency. But here, as in thunder storms, no definite connection has yet been shown between the various suggested agencies and the actual processes of the storm winds; while in all cases, the action of convection is in accord both with the conditions in which local storms occur and with the processes that they exhibit.

269. The vortex of tornadoes. The development of a whirling motion has already been shown to be a necessary feature in the growth of violent cyclones; a simple radial convectional inflow is unable alone to cause winds of destructive strength. It is the same with tornadoes. Their destructive winds are not radial inflows; but they become violent only when a vorticular whirl is developed.

In this connection, it is important to distinguish between bodily rotation, such as that of a wheel or of the earth, and vorticular whirling, as in a water eddy or atmospheric vortex. In the former, all parts rotate at a constant angular velocity about the central axis, and the linear velocity increases with the distance from the axis. In the latter, the angular and the linear velocity rapidly increase *towards* the axis. It is for this reason that the tornado vortex is dangerous only at a small distance around the funnel cloud.

In nearly all cases where the direction of tornado whirling has been determined in this country, either by observation of the funnel cloud or by the distribution of objects overturned or blown about by the whirling winds, it has been found to be from right to left; that is, in the same direction as that of the cyclonic spirals of this hemisphere. It is possible that in some cases the direction of turning may depend on the accident of greater strength of inflow on one side than on the other; but as a rule, the systematic turning from right

to left is manifestly dependent on a constant cause, such as the earth's rotation. It cannot, however, be supposed that the area, from which the inflow comes at the base of a tornado, is large enough to introduce a determining deflective effect from the earth's rotation; and therefore the prevailing direction of tornado whirls should be regarded as determined by the vorticular movement of the cyclonic winds around the center of low pressure; these having been previously determined by the earth's rotation. It is a general mechanical principle that when a small whirl springs up in a larger whirl, the two must turn in the same direction. It is for this reason that our cyclones turn in the same direction as the whirl of the circumpolar winds. The same principle explains the agreement in the direction of rotation and revolution of the planets around the sun and of the moon around the earth. Indeed, all these larger and smaller turnings, from the greatest to the least, must be regarded as the continued inheritance from the original impulse by which the rotation of all the bodies in the solar system, including the sun, was determined.

270. Ferrel's theory of tornadoes therefore begins with the occurrence of an especially active convectional ascending current within a thunder storm; the possibility of such currents being indicated by the greater activity of ascent in some thunder heads than in others; while the actual occurrence of active ascending currents in tornadoes is indicated by the vertical component of the whirling winds which lifts heavy objects high above the earth. The ascending current draws on the lower warm and moist air for its supply; but as the air makes part of a large slowly-whirling cyclonic storm, the tornado inflow must develop a central whirling vortex, which shall turn in the same direction as the parental cyclone. As the inflow is drawn in from the margin towards the axis, slowly ascending at the same time, the whirling component of its velocity is greatly increased, and when near the center, it may attain an irresistible violence. At a moderate distance above the ground, perhaps a few hundred feet, the effect of friction is so small that the inflow becomes almost a perfectly circular whirl of extremely high velocity close around the axis, forming a central core of low pressure, very similar to that of the eye of a tropical cyclone. But the lower air is prevented by friction with the ground from attaining so great a whirling velocity; its centrifugal force is therefore less than that of the stronger whirl above it; and it is consequently drawn rapidly into the overhanging core of low pressure. It is therefore believed that the spiral inrush of the lower air into the low-pressure core made by the higher whirl constitutes the destructive blast of the tornado.

The student should guard against forming too rigid a conception of this theoretical process. The activity of the ascending current and the violence of the whirling inflow must vary from time to time; the axis of the tornado need not be vertical, but may incline somewhat to one side or another; it need not

be a straight line, but may slowly twist and writhe, after the fashion of the empty axial core of water eddies, easily observed; the low pressure of the core must vary with the strength of the lateral inflows, which cannot be of uniform value. When allowance is made for the natural irregularities by which the ideal process is complicated, it may be fairly claimed that the convectional theory of tornadoes gives a reasonable explanation of all the phenomena that have been observed in these storms.

271. Central low pressure of tornadoes. The inferred low pressure of the tornado core has never been determined by observation, and probably never can be; but analogy with other whirls renders its occurrence highly probable. The circumpolar whirl of the terrestrial winds has been found competent to reverse the expected high pressure of the cold polar regions into low pressure; and the stronger whirl around the south pole causes a lower pressure than that which is produced by the slower whirl around the north pole; indeed, if it were not for the resistances caused by the intermixture of upper and lower currents in cyclones and anticyclones, the polar pressures might be much lower than they now are. Again, the central low pressure of tropical cyclones, initiated by high temperature, is greatly intensified by the development of centrifugal force in its whirling winds. It is therefore reasonable to believe that the excessively rapid whirl of the winds in the tornado vortex close around the axis must develop a vastly greater centrifugal force than occurs even in tropical cyclones; and the occurrence of low pressure within such a vortex cannot be doubted. Various facts confirm this expectation. A number of examples have been reported in which the walls of buildings have been blown outward, as if by the explosive expansion of the inside air, when the low pressure of the tornado core passed overhead. It is possible that other explanations may be offered for this peculiar fact; but none appear so reasonable as this one. For example, a critical observer, describing the destruction of a factory in a tornado at Arlington (then West Cambridge), Mass., in 1852, stated: — "The whole effect produced, and to my own mind well and clearly defined, was precisely what we should have if we could suddenly place in a vacuum a building filled with atmospheric air of ordinary tension. Even the foundation walls were inclined outwards, and there was every evidence of a force acting from the interior to the exterior." In some tornadoes, it is reported that corks have been drawn from empty bottles, as if by the expansion of the air from within.

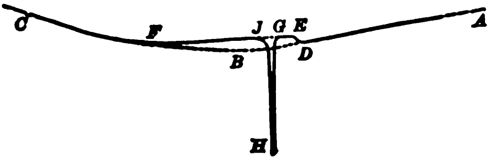

Fig. 103.

The atmospheric pressure observed along an east and west line, south of the center of an ordinary cyclonic

storm, may be indicated by a concave curve, as *ABC*, Fig. 103. If a thunder storm occurs about the time of lowest pressure — that is, near the western border of the southerly winds — it causes a slight rise of the barometer, not from increased weight, but from the downward reaction of the rapidly expanding and ascending mass of air aloft; this is represented by the modified curve, *ADEFC*. Now if a tornado occurs in such a thunder storm, its whirling vortex causes an extremely low pressure within a slender cylindrical core; and the previous curve would become *ADEGHJFC*. The changes of pressure described for the cyclonic storm and for the thunder storm are matters of ordinary observation; those drawn for the tornado are matters of reasonable inference.

272. The tornado funnel cloud. Accounts of tornadoes and water spouts frequently mention the descent of their funnels from the heavy overhanging clouds, as if there were actually some descending motion; but there is good reason for believing that the descent is only a deceptive appearance. While it is generally true that the movement of clouds indicates the movement of the air in which they are formed, this is not always the case. The base of an ordinary cumulus cloud is fixed at a certain height, although the air is constantly rising through it. Stationary clouds stretching out from mountain peaks, or standing in fixed waves, while the winds in which the clouds are formed are moving onward, convince us that the outline of a cloud merely marks the limit of a space within which the vapor of the air is condensed into visible cloud particles. The lower front edge of thunder storm clouds may often be seen growing to windward; the eastward advance of the space within which the ascending air is cooled by expansion being more rapid than the westward ascent of the inflowing wind. The tornado funnel cloud is even a more striking example of this contradiction between apparent and real motion. The funnel seems to descend, because, as Franklin clearly said in 1753, the moisture is condensed "faster in a right line downward than the vapors [cloud particles] themselves can climb in a spiral line upwards."

This may be explained by Fig. 104. Suppose the observer is looking north towards the funnel *EF*. Consider now the condition of the air at *A*, at a moderate height above a point, *B*, which lies several hundred feet southwest of the vortex. The temperature and humidity of the air at *A* is such that if it ascends vertically, it would become cloudy at the height, *H*, in the base of the great overhanging thunder cloud. Instead of rising vertically, the air from *A* moves along an inflowing and gradually ascending spiral path, *ADC*, towards the lower pressure of the tornado core. On reaching the point *C*, it is cooled both by expansion due to ascent, *AC'*, and by expansion into the low pressure core; and if at *C* the cooling due to expansion into the low pressure of the vortex equals the cooling which would be produced by ascend-

ing through the additional height, *C'H*, the inflowing whirling air will become cloudy.

The combination of these two causes of cooling gives full explanation of the form of the funnel cloud. It first appears as a somewhat depressed portion of the overhanging cloud mass; and from this it is inferred that the whirl of the tornado begins high above the surface of the earth and extends its action downward, as might have been expected from the probable form of the vertical temperature gradient of strong thunder storms, as given in Fig. 99, where the flatter part of the line, *DG*, locates the place of greatest instability at a considerable height above the ground. As the ascent and inflow of

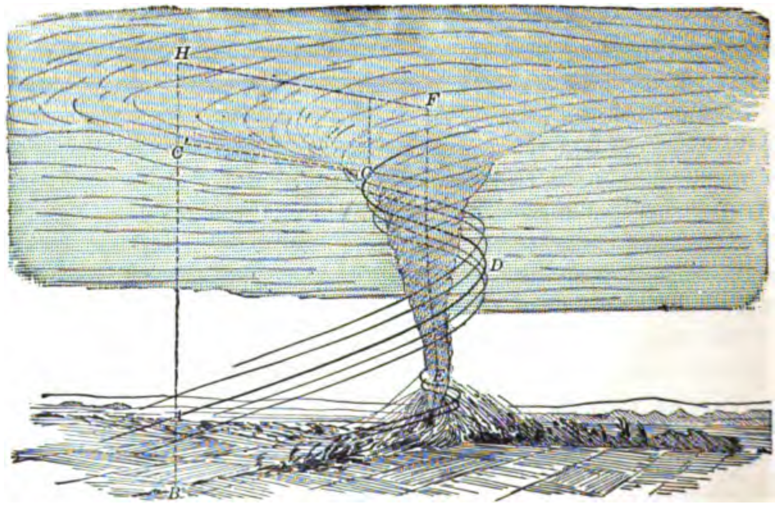

FIG. 104.

the growing tornado become stronger, its whirl is strengthened, and the amount of cooling due to expansion into the axial core increases; hence the funnel cloud forms at lower and lower levels. Finally, in tornadoes of full strength, the funnel seems to reach down to the ground, and in the language of ordinary descriptions, "it destroys everything that it strikes." This should be interpreted to mean that if the violence of the vortex is sufficient to cause cloudy condensation close down to the ground, it must also be strong enough to destroy everything in its path. The lower part of the funnel is of less diameter than above; because a closer approach must be there made to the hollow core before cloudiness begins. Sometimes the lowest part of the funnel widens, presumably from the gathering of dusty rubbish from the ground.

There is frequent mention made in the descriptions of tornados of two clouds, one darker than the other, which rush together and form the destructive whirl. Successive observers of the same tornado make the same

report; and hence it must be supposed that this process is a continuous one. In such cases, it is not the rushing together of the clouds that causes the destructive wind, but the rushing in and whirling around of the wind that continually creates the clouds and carries them inwards towards the vortex.

After lasting half an hour or an hour, and leaving a path of greater or less destruction across the country, the tornado weakens, as if the supply of exceptionally warm and moist air on which its life depends were then exhausted, or as if the upward path of escape were deformed and confused. The funnel gradually withdraws from the ground, disappearing in the clouds above, and the storm is over. But when one tornado is formed, others often appear in the same neighborhood, and thus a succession of whirls may traverse the country, as in the example of February 19, 1884; although the number of tornadoes then reported is altogether exceptional.

273. The progression of tornadoes. Tornadoes in this country generally move easterly or northeasterly, sometimes southeasterly; seldom in other directions. Their velocity of progression commonly varies from twenty to forty miles an hour. As the vortex is small and its progression rapid, less than a minute suffices to carry it past any given point. The duration of tornadoes ranges from half an hour to an hour or more; hence their length of path may reach thirty to fifty or more miles. Examples in which a much greater length of path is reported probably consist in reality of two or more whirls, forming successively almost in the same line. Although occurring within the area of southerly surface winds, the advance of tornadoes is more accordant with the direction of the higher currents; hence it may be inferred that the place of escape of the ascending currents is born along within the great nimbus clouds by the westerly overflowing winds.

Instances were mentioned in Section 257 of local thunder storms that stood distinctly within the area of southerly winds and that were followed as well as preceded by high temperatures. It is not yet clear from reported records whether this is generally or occasionally the case with tornadoes also. It is frequently stated that the warm, sultry weather which preceded a tornado was followed by cooler weather; but it is not yet made certain that tornadoes always occur close along the belt of separation between the southerly and westerly winds. Special observation might well be directed to this question.

274. Protection from tornadoes. The approach of tornadoes is so rapid and their arrival follows so soon after their first appearance that there is very seldom time for those who happen to be on their path to escape their fury. If a tornado is seen approaching obliquely, so that the continuation of its path will carry it to one side of the observer, but not far distant, he should lose no time in running away from it; and if a distance of five hundred feet

is gained before its arrival, serious danger is pretty surely avoided. If the tornado seems to be coming directly towards the observer, it is safer to run to the northern side of its path; for on that side its whirling winds, blowing to the west, are weakened by the progressive velocity of the whirl which carries it to the east. For this reason, the path of the vortex does not lie along the middle of the path of destructive action, but somewhat north of the middle.

In case of the sudden approach of a tornado unseen, as at night, when its coming is only known by the roaring of its winds, the southwest corner of a house cellar is regarded as the safest place of refuge. Sometimes special underground cellars are prepared beforehand for case of need; and these are provided with a bar, an axe and a saw by the more cautious, in order to assure means of escape in case the house is demolished overhead.

Although individual tornadoes are excessively destructive to everything that lies in their path, yet their limits of action are so narrow and our country is so wide that the danger from tornadoes at any one place is much less than has been supposed. The annual loss to the country by fire and flood greatly exceeds that caused by tornadoes.

275. Observation of tornadoes. The rapid passage of a roaring tornado is not a time when deliberate observation of its action can be expected. Yet if a person happens to stand a few hundred yards on one side of its path, a close scrutiny of the funnel and the clouds overhead might be safely made. If such a person had in mind the interpretation of tornado action as here presented essentially in accordance with Ferrel's theory, his attention might be critically directed to such features of the storm as need examination in repeated occurrence. He should examine the funnel to determine the character of its rotation; he should look closely to discover the movements of cloud wisps or of trees and other objects in the funnel; he should examine the relation of movements in the funnel to those in the greater overhanging clouds; the behavior of the two clouds, whose rushing together is so often mentioned; the occurrence of descending clouds sometimes noted at some distance to one side of the vortex. In a larger way, the relation of the tornado to the fall of rain or hail from the thunder cloud should be examined, for heavy precipitation commonly occurs at a moderate distance from the tornado, rather than immediately around its vortex; the distribution of temperatures and the relation of electric action to the funnel also need attention, as different accounts vary greatly in these matters.

After the passage of the tornado, the peculiar effects of its destructive winds should be recorded: a map should be prepared to show the place of buildings destroyed and of trees overturned, and the path of objects carried by the winds. That tornadoes are destructive is only too well known, but the particular method of their action as indicated in their effects should be care-

fully determined. As a rule, the evidence of vorticular action in the attitudes of overturned trees and in the distribution of fragments of buildings is not so clear as might be expected from the visible whirling of the winds in the funnel cloud. Greater or less confusion results in the first place from the increased strength of the indraft on the southerly side of the whirl, as already explained; and in the second place because the destruction of objects is not accomplished all at once, but irregularly and successively, according to their resistance and to the storm's strength. For example, on the northern or left side of the track, trees are sometimes blown down almost in a forward direction, as if by the rear inflow; and therefore the backward whirl of the winds on this side is more or less confused with forward action. Yet sometimes a field of grain may record a partial circuit of the winds, as if they had suddenly fallen on it, and prostrated all the stalks at once in a sweeping curve, with a radius of several hundred feet. The indications of whirling motion found in prostrated trees is generally as follows: trees blown down backward are found only on the northern side of the track of the vortex; they are often crossed over by other trees falling to the southeast or east, as if by the later winds in the rear of the whirl. On the south of the vortex, many trees are laid nearly parallel with the track; but those first blown down turn more to the north, and those last overthrown turn more to the south. Sometimes the distribution of identifiable fragments from buildings gives indication of a curved course through the air: at the Lawrence, Mass., tornado of July, 1890, the southern windows of a house south of the track were broken by rubbish carried from other houses several hundred feet to the northwest. Such facts as these should be written down on the ground, in order that no mistakes of memory may occur.

276. Waterspouts. The behavior of waterspouts at sea is so closely like that of the spouts caused by tornadoes when they cross rivers or ponds that there can be no doubt of the essential similarity of these vorticular storms on land and sea. Waterspouts possess a tapering funnel cloud, first seen as a small pendant from the under surface of the overhanging clouds; then apparently descending to sea level, where the greatly agitated waters rise to meet it. Although these spouts seem to draw water up from the sea, they consist of fresh water for the greater part; and hence must be regarded as the product of vapor condensed from the air. There is an old account of a vessel on which a waterspout fell. A flood of water poured on the master, so that he was obliged to lay hold of what was nearest to him to escape being washed overboard. He was asked afterwards if he had tasted the water. "Taste it," said he, "I could not help tasting it; it ran into my mouth, nose, eyes and ears!" "Was it, then, fresh or salt?" "As fresh," said the captain, "as ever I tasted spring water in my life."

Waterspouts seem to be most common in the warmer and calmer seas; but they are also recorded in middle or higher temperate latitudes and in the presence of moderate winds; and a good number of examples have been recorded on the Gulf Stream in winter, in the presence of cold westerly winds blowing off the colder land. In some of these cases, the instability on which the spouts depend may arise from local causes; but in others, and particularly in the last mentioned, the instability appears to depend on the importation of air with a temperature much lower than that of the water over which it advances, much as was the case with tornadoes on land.

A few records have been made of the appearance of descending currents within the core of a waterspout; from which it must be concluded that the interior and exterior parts of the spout have opposite motions. It has been suggested that these descending central currents are streams of rain, falling from above into the nearly empty core within the whirling spout; and that such currents are more likely to characterize waterspouts, where the lower inflow close to the sea surface is little retarded by friction, and hence cannot enter the core easily; while in tornadoes on land, descending currents would seem to be less likely to occur, because their place is taken by inflow and ascent of the surface currents, whose centrifugal force is not sufficient to hold them out of the low pressure close to the axis. The reported descending currents in waterspouts may therefore be compared with the inferred descending currents in the eye of tropical cyclones at sea; while the presence of only ascending currents in tornadoes may be compared with the inferred ascending currents about the usually cloudy center of cyclonic storms in our latitudes and on land. Closer observation, with suggestions of this kind in mind, may some day determine these minor points.

CHAPTER XII.

THE CAUSES AND DISTRIBUTION OF RAINFALL.

277. Causes of rainfall. When vapor is condensed in sufficient quantity, it falls from the clouds and reaches the earth as rain or snow. All forms of atmospheric precipitation are included under the general term, rainfall.

The occurrence of rain or of its winter equivalent, snow, is in nearly all cases associated with overgrown clouds; and in the temperate zone at least, these are usually products of cyclonic or of local storms. It occasionally happens that rain or snow falls from the clear sky; but this is highly exceptional, and its amount is small. The processes which produce clouds may always, if carried far enough, bring forth rainfall. Mention has already been made of this in the case of the great overgrown cumulus clouds of warm summer weather. The small clouds of morning are succeeded by greater ones towards noon, and these by even more massive clouds two or three hours later. Such cloud masses may be many miles long and wide, and at least four or six miles high, drifting along in the high-level currents of the atmosphere and yielding heavy rain to the earth below. Cyclonic cloud masses are many times larger.

The cause of rainfall is not far to seek. Every cloud particle serves as a center for additional condensation in the cooling saturated air. The particles must be of slightly unequal size; the larger ones are not so easily borne up in the air as the smaller ones are; collisions must occur, and when two drops coalesce, the resulting larger drop tends to fall more rapidly than either drop fell before. In the vertical or obliquely ascending currents, in which clouds and rain are so commonly formed, the drops may grow to a size large enough to cause them to fall through the rising air. As the temperature of the drops is then lower than that of the damp air through which they fall, continuous condensation is provoked on their cool surfaces. Collisions will be more frequent than before, and the drops will grow more and more rapidly as they fall to the base of the cloud, where their largest size is reached: $\frac{1}{50}$ to $\frac{1}{25}$ of an inch in fine rain; $\frac{1}{10}$ of an inch or more in the heavy pattering rains of summer. On falling below the cloud into non-saturated air, the size of the drops decreases by evaporation. In dry regions or seasons, it is not uncommon to see a trail of rain falling from the base of a lofty rain cloud, and entirely disappearing before reaching the earth. Such rain clouds may be seen rising over the ridges of our Rocky Mountains, bringing dark streams of rain along beneath them; and yet only a few large drops reach the thirsty ground; but when the lower air is damp, as under winter cloud sheets, the rain that falls from the cloud for the most part reaches the earth. Sometimes the

rain drops of winter storms are frozen into clear ice pellets, when falling from a warm upper current into cold surface air; this should be called frozen rain, and not hail, which is of quite different form and associations, as is explained below.

278. Snow. In case the process of condensation occurs at temperatures below 32°, the vapor then crystallizes as it condenses and forms snow flakes or ice needles of varied forms, but always presenting angles of 60° and 120°, characteristic of crystallized water. In quiet snow falls, the crystals are remarkably perfect and by far the greater number of them are of one form. As each crystal is developed in this case from a single center, their growth must be explained by continuous condensation on some initial nucleus, and not by collision, such as presumably occurs in the formation of faster falling rain drops, or in the matted flakes of windy snow storms. In milder winter weather, when snow falls into a warmer surface layer of air, it partly melts and reaches the ground as sleet. A more complete melting would deliver it as rain; and it is probable that most of our winter rain has had the form of snow while it was still in the higher clouds. Indeed, a part of our summer rains also may be formed as snow in the upper parts of the thunder storm clouds; and if caught on high mountains, the snowy form is preserved. When precipitation occurs in the polar regions at temperatures lower than −5° or −10°, small ice needles and not snow flakes are formed.

279. Hail consists of compacted ice and snow, often arranged in roughly concentric layers, taking the form of little pellets or balls, commonly called hail-stones. It does not fall in winter, but is a common accompaniment of thunder storms, even though they occur chiefly in the warmer regions and seasons of the globe. Occasionally hail-stones show a crystalline structure. They are sometimes of remarkable size, up to several inches in diameter; and they may then cause serious destruction to trees, crops and buildings. Hail should be distinguished from the clear icy pellets of frozen rain, which sometimes fall in the winter season, but never in the summer; and also from round pellets of snow of loose structure, called soft hail, again a product of winter instead of summer.

The association of hail with active convectional storms in the warm season suggests that it is produced by the freezing of rain drops that have been formed at low levels, and that are then carried upwards by the central ascending currents to altitudes where the temperature is very low; there they become coated with a layer of snow, increasing in size until they fall through the less active currents near the margin of the storm; still increasing in size by further condensation on their cold surfaces during descent through the clouds; and perhaps again carried inward at the base of the cloud and upward once

more through its center; until at last they become too heavy for further carriage and fall to the ground. Before their fall, a curious rattling may sometimes be heard in the air, as if caused by their noisy collisions. As a general rule, hail is larger and more plentiful in violent than in moderate thunder storms.

Although associated with electrical storms, there is no sufficient reason for regarding electricity as the chief agent in the production of hail. It is true that hail-stones may still be so distinctly electrified on reaching the ground that they may leap again a few inches into the air; not only by elastic rebound, but as if by electric repulsion. Yet as with the other products of thunder storms, the electric condition of hail-stones seems to be essentially a result of the processes of their formation, and not primarily a cause of their formation. The supposition that they rise and fall by electric attraction and repulsion between the two cloud layers of which thunder storms have been said to consist does not find support in the present knowledge of the structure of thunder storms, or in the estimates of the force that electrical attraction and repulsion could attain in the atmosphere.

Like the rain of thunder storms, their hailfall is distributed in belts, whose breadth depends on the size of the storm, and whose length depends on its duration and velocity of progression. In the larger linear thunder storms, hail seems to fall only from certain parts where the storm is of great violence.

280. Conditions of rainfall. There can be no question that ascensional movements in the atmosphere of whatever cause are the most effective means of condensing vapor so plentifully and rapidly as to produce rain: hence the greater rainfall of regions frequented by cyclonic storms or thunder storms, where the air is given a vertical component in its movement; hence also the greater rainfall of mountains, especially on their windward slopes, where the air is forced to ascend in crossing them. There are, however, two other processes of cooling already mentioned in connection with the development of cyclonic and other clouds, that should be referred to here again. The first of these is the movement of masses of air poleward so that their attitude with respect to sunshine is changed, and radiation coming to be in excess causes their temperature to fall. These currents may become cloudy in so great a mass as to yield rain: hence our southerly cyclonic winds or moist siroccos are not only generally cloudy but frequently rainy as well. In this country, it is as a rule only cyclonic winds that move poleward directly and rapidly enough to produce speedy condensation in this way.

A second process deserving mention is that which appears when a current of moist air from over the ocean advances upon a cold winter land. Its mass may then become cooled not simply by conduction, which extends but a little

way above the earth, but by radiation from the air, especially from the cloudy air, to the ground. Landward winds may thus become cloudy in large volume, even to the point of yielding rain; but when rain appears, the winds are generally in this case also found to be within the influence of a cyclonic storm, in which, as has just been stated, some vertical motion aids the other causes of cooling. Rainfall as a result of the mixture of two masses of saturated air at different temperatures does not seem to be common.

The exceptional occurrence of rain falling from a clear sky is called *serein*. It is rarely noted and is not well understood.

281. Cooling caused by rainfall. Distinction must be carefully made between the adiabatic increase of temperature by compression in descending currents of air, and the maintenance of a nearly constant temperature in falling rain drops or snow flakes, which cannot be compressed. While the descent of a current of air from the height of a thunder storm summit would give it a high temperature at sea level, the descent of rain drops or snow flakes from similar heights causes a distinct cooling of the warmer lower air through which they fall. A considerable part of the lowering of temperature that is commonly noted during a summer rainfall must be ascribed to this process.

282. Variation of rainfall with altitude. Clouds attain their greatest frequency at a moderate height above the earth, averaging perhaps a half mile or a mile, because the dew-point to which the air is cooled by the various processes of cloud-making is most commonly encountered at this altitude; their greatest density is found at about the same height because condensation at greater altitudes and hence at lower temperatures is attended by less plentiful exclusion of vapor. For the same reason, the greatest measure of rainfall in a given region is not at sea level or at the level of the earth's surface, but at a certain moderate altitude in the atmosphere, varying with the region and the season. At lower levels, some of the precipitation falls from the clouds into non-saturated air and is redissolved; at greater altitudes, less precipitation is formed. In the Alps, the maximum is found at about 3000 or 4000 feet in winter; but in summer, when the air is relatively drier, the maximum appears to be at a greater height than the records reach. In the southern ranges of the Himalaya, the level of maximum summer rainfall is at 4000 feet; and it is near this height that Cherrapunji lies on the Khasia hills, where the heaviest rainfall of the world is found (Sect. 301): in winter the maximum rainfall is at about 20,000 feet. In this country, an increase of rainfall up to a certain height undoubtedly prevails, especially in the western territory, where the air is generally so dry; but records are not yet made to determine it. The rainfall maps of our western area, based

on observations made for the most part on the plains and lowlands, probably do not represent the full value of the precipitation of that mountainous region. The lofty upland surface of the Aquarius plateau in southern Utah is well watered and bears an extended forest, while the lower plateau country about it is a desert; but no sufficient indication of this contrast can at present be given on the charts, on account of the absence of high-level records.

In Europe, where records of rainfall have been more generally maintained on highlands and lowlands, the rainfall charts correspond to a remarkable degree with the relief of the country. The highlands of the northern Atlantic coast are shaded dark for a heavy rainfall; the Pyrenees possess a plentiful rainfall between the dry lowlands of southern France and the semi-arid plateau of northern Spain. The Alps form a center of strong precipitation. The small rainfalls of the basins of Bohemia and Hungary are surrounded by heavier rainfalls on the enclosing mountains of Germany and Austria. The mountains of the Caucasus have a heavy rainfall between the dry steppes of southern Russia and the arid plateaus of Asia Minor. The more accurate the charts, the closer this relation appears to be.

The rainfall on mountain ranges is greater on their windward slopes. Little difference in this respect is perceptible on ranges so low as our Appalachians, and on which the rainy winds blow from different sides. The Sierra Nevada has more rainfall on its long western slope than on its precipitous eastern descent. The Pyrenees are watered chiefly on their northern side; the Himalaya on their southern. The equatorial Andes have heavy rains on their eastern slopes; the Chilean Andes receive rain chiefly on the western slope.

283. Measurement of rainfall. It is desired to measure the depth of the sheet of water that would lie on level ground after a rain if none of the water were lost by evaporation or by soaking into the soil. This is done by exposing a cylindrical vessel or rain gauge to the storm and measuring the depth of rain or snow that it receives. A good gauge should have a truly circular rim; a diameter of at least five or six inches being recommended. The edge should be sharp, with a vertical face on the inside. The gauge should be placed in a level and open field, removed if possible from all trees and buildings by at least twice their height; it should be fastened in position, to avoid overturning by the wind. The rim should stand a foot above the ground; it should be carefully levelled. Once placed in a well-selected situation, subsequent change should be carefully avoided; but if required, a full account of the change should be entered in the record book. In order to avoid loss by evaporation, a movable funnel is generally placed within the gauge, thus protecting the water that lies beneath it from loss to the air.

The measurement of the amount of rain collected is best done by pouring the water from the gauge into a measuring tube of a certain smaller diameter, so that its area shall be one tenth of that of the gauge. The water then rises in the tube to ten times the true depth of the rainfall. This magnified depth is then measured by a graduated stick, the record being made to a hundredth of an inch. Record should be made, if possible, at the close of every storm and always once a day; although some observers measure the rainfall only at a certain hour every day, without regard to the time when the rainfall ceased. The amount measured should always be entered in the record book before the measuring tube is emptied.

The easy drifting of light snow makes its measurement a matter of much uncertainty. It can seldom be correctly determined by the amount that is caught in gauges, unless the wind has been very light. It is recommended that observers measure the amount of snow lying on the ground in open woods, where drifting is slight. The measure may be made with a stick while the snow is on the ground; or a section of snow may be cut by the rim of the gauge, and the amount of snow thus secured may be melted and measured as rain. Melting is best done by adding a measured amount of warm water. Light snow is generally eight or ten times as deep as the corresponding amount of rain.

Records of rain and snow from gauges on buildings in cities are, as a rule, defective, because of the eddies of wind by which too much or too little rain is carried into the gauge. Such measures may serve to indicate generally whether the fall is light or heavy, as is required in weather reports; but they should not be accepted for the climatic tables of a district. Self-registering rain gauges have been devised, by which the fall is recorded every five minutes, but these are seldom employed. No satisfactory means have been devised to measure the rainfall at sea: only the time of occurrence, the relative frequency and the estimated amount of rainfall are reported in marine observations.

284. Records of rainfall. The data desired in this connection are: the amount of precipitation in every separate fall, but when brief showers follow one another, the whole fall may be measured at once; the time of beginning and ending; and if possible, the direction of the wind at these times. The rainfall of every day should be determined late in the evening during a long storm. When the last day of a month is rainy, the measurement should be delayed till as near midnight as possible, in order to give a correct monthly total. The records of rainfall are generally summarized as follows: Total rainfall for each month; snow on the ground on the 15th and at end of month; share of monthly fall in the form of snow; dates of first and last snowfall; number of rainy days; that is, of days on which more than one hundredth of

an inch of rain or snow fell; maximum daily fall of the year. After the records have been carried on over a period of years, the normal or mean monthly and annual rainfalls should be determined. In our country, at least ten years record is needed before the mean annual total can be determined with acceptable accuracy; and a thirty years record is needed for the monthly means, as the fluctuation in their values is often great. The maximum fall for a given month may be many fold greater than the minimum. The plus or minus departure of the fall of each month from its normal should be stated. The average number of days in each month on which a hundredth or a tenth of an inch falls, or the probability of rainy days, constitutes an important climatic factor. It is desirable to determine the average number of rainy spells for each month and the average fall in each spell. The share of rainfall supplied by winds or storms of different character should receive attention; and the summary for the year should subdivide the total as far as possible according to origin.

No regions where continuous records have been maintained are found to be absolutely rainless. Southeastern California and western Arizona have certain stations where the mean annual rainfall is under two inches, and where in single years less than an inch has been gauged. Other desert regions, such as central Arabia or the interior of the Sahara may have even less, but records have not been kept there to make this certain. The greatest annual rainfall is found in India and the East Indies, where many stations record over a hundred inches a year. The most remarkable of these is Cherrapunji, at an elevation of 4,455 feet on the southern slope of a subordinate range of the Himalaya mountains, north of the head of the Bay of Bengal, with an average annual rainfall of 474 inches, of which over 400 fall in the five months from May to September, or during the summer monsoon. A fall of 40.8 has been measured at this station in a single day (June 14, 1876), and over 600 inches or fifty feet have been collected in certain years. The average fall for the five rainy months is almost three inches a day. Here truly "it never rains but it pours." The rainfall of Mahableshwar on the bold western slope of southern India at an altitude of 4,540 feet is hardly less remarkable, reaching an average of 261 inches, of which 251 fall from June to September inclusive. It is noteworthy that at a little distance east of this station on the relatively even plateau of the Deccan, the rainfall is reduced to less than 20 inches a year.

The greatest mean annual rainfall of this country is 101.87 at Neah Bay, Wash. (or including Alaska, 111.72 at Sitka), and the greatest single annual fall is 123.23 at Neah Bay in 1886 (or 140.26 at Sitka in 1886). The following items have a statistical interest in this connection:[1] Excessive monthly rainfalls: at Upper Mattole, California, January, 1888, 41.63; at Alexandria,

[1] See Greeley's *American Weather*, in which many facts of this kind are collected.

Louisiana, June, 1886, 36.9. Excessive daily falls amounting to ten or twelve inches have been recorded at several stations. Brief downpours: at Washington, D. C., June 27, 1881, 2.34 in 37 minutes; at Philadelphia, July 26, 1887, 0.62 in seven minutes. Much heavier falls have undoubtedly occurred in cloud bursts, such as happen in the western states and territories, but no precise measure of their amount is at hand. The amount of rainfall in single storms is sometimes excessive over a large area of country, producing disastrous floods. On February 11–13, 1886, more than five inches were recorded over an area of 5,000 square miles in southeastern New England; the Johnstown flood in Pennyslvania, May 30–June 1, 1889, was estimated at over eight inches upon an area of about 12,000 square miles, with somewhat less fall on a much larger surrounding area, causing a terrible destruction of life and property. In northern India, September 17–18, 1880, ten inches of rain fell over an area of 10,000 square miles; its weight being 7,248,000,000 tons. It is manifest that the liberation of latent heat from so vast an amount of rain — or the release of so great a supply of stored solar energy — must be the means of doing an enormous amount of work.

285. Relation of rainfall and agriculture. When the annual rainfall is under eighteen inches, agriculture can seldom be safely practised without irrigation. Grazing then becomes the chief occupation, as is now the case over a large extent of our western plains, between the 98° meridian and the Rocky Mountains; and the people return in a measure to the roving life characteristic of the aboriginal inhabitants of semi-arid regions. When the rainfall is less than twelve inches a year, the region is reduced to a desert, and the water supply is too small to be of service in irrigation, unless in small areas, or on the banks of large rivers. On the other hand, the tropical regions where the rainfall rises above a hundred inches a year are so luxuriantly overgrown as to make their occupation a difficult matter. The general rainfall of the eastern part of our country or of western Europe, with an annual total varying from forty to eighty inches equably distributed through the year, is an amount under which human occupations are best developed.

The distribution of rainfall through the year is a matter of great moment. The northeastern part of the United States is favored in having the average value of the precipitation in successive months comparatively equable; droughts are exceptional, but when occurring are found in one season about as frequently as another. In Florida, the summers are wet and the winters are comparatively dry. In eastern Nebraska, there is a similar distribution of rainfall from dry winters to wet summers. In California, the reverse is true; the summers have a continuous drought, making the ground dry and dusty; the winters are cloudy and wet. These variations will be found to depend chiefly on the system of the general winds.

The irregularity in the monthly rainfall is an important matter, especially in regions where the annual supply is moderate. Even in the well-watered states east of the Mississippi, a deficiency in the summer rainfall sometimes causes droughts; for example, the southern Atlantic states had in September, 1888, 1.84 inches; while the same month of the preceding year had 9.47, or over five times as much: this variation appears to result chiefly from the variation in number, paths, and activity of cyclones in different years. On the margin of the western Plains, where the total rainfall is hardly enough for agriculture, the departures from the normal monthly fall are of more serious import: a succession of rainy seasons tempts settlers further and further west, and when a series of drier years follows, the distress occasioned by the failure of crops becomes a calamity.

In India, where the year is divided into three seasons, the cold, the hot, and the wet seasons, the crops are preserved in the dry season by irrigating canals fed from rivers rising in the mountains, or in a smaller way by water pumped up from the rivers; if the water in the rivers in insufficient, or if the rains arrive late or are deficient, famine results, and the people of that great country die by the thousands. In more recent years, with the improvement of the irrigating canals, and with the better means of transportation of the plenty of one province to supply the need of another, the danger from this source is lessened.

286. Snowfall. In regions where the winter snowfall covers the ground to a thickness of a foot or more and remains unmelted for a considerable period, it exercises an important influence on the temperature of the air. Being of relatively loose texture, it is a poor conductor and thus very effectively prevents the escape of heat from the ground; at the same time, the surface of the snow, losing its own heat by radiation and absorbing very little insolation, falls to a low temperature, and thus cools the air lying upon it. The presence of a heavy snow-cover during winter thus protects the ground from deep freezing and at the same time determines the occurrence of very low temperatures in the air.

The opening of spring is much delayed in regions where the winter snows accumulate to a considerable depth; for until all the snow is melted, the surface cannot gain a temperature above 32°. The low temperatures of the polar regions, where ice and snow prevail and last through the brief solstitial season with its high values of insolation, have already been explained in this way (Sect. 88). Near our eastern coast, where rain and snow rapidly succeed one another in winter time, it frequently happens that a heavy snowfall will be almost entirely melted a few days later by a mild rain, and thus the precipitation of two storms will be delivered quickly to the streams. Our floods of winter and spring are chiefly caused in this way.

Lofty mountains in all latitudes and plateaus in the polar regions often receive a greater supply of snow in the cold season than is melted in the following milder season. The thickness of the snow cover then continually increases until a movement towards lower ground is established to dispose of the excessive supply; if lying on steep mountain slopes, the snow falls in avalanches into the adjacent ravines, filling them to a great depth. With increasing pressure and especially with the aid of percolating water from surface melting and from occasional rain storms, the accumulated snow is gradually welded into ice. The mass thus formed slowly creeps downward, following the slope of the ground, and enters lower and milder levels as a glacier; finally ending when the melting of its extremity balances the supply from downward creeping; or in the polar regions, ending in the sea where its margin breaks off and forms icebergs, which are then borne away by currents.

The value of snow as a store of winter precipitation for summer use is very great in arid regions; and in the coming century we may expect to see the water of spring freshets that now runs to waste from our western mountains utilized in large part by detaining it in reservoirs in the upper narrow valleys, and leading it out along the valley sides when desired in artificial canals until it reaches the flat divides of the interstream surfaces on the arid plains, where it can then be distributed over large areas. This is already done in a small way, but such works will have to be greatly increased as the need for them becomes more and more pressing.

287. Ice storms. Regions of strongly variable temperature are subject to occasional winter storms in which the precipitation occurs as rain, but freezes as soon as it touches any solid body, such as the branches of trees, or telegraph wires, or the ground. This happens when the ground and the lower air have been made excessively cold during a spell of clear anticyclonic weather, when a moist upper current in advance of an approaching cyclone brings clouds and rain (Sect. 245). Serious damage is caused by breaking down over-weighted wires and branches at such times. Wires may be increased in weight ten or twenty fold; and twigs even more than a hundred fold. New England is particularly subject to such storms, although they are not by any means common; in the winter of 1886, three ice storms occurred in January and February, but this was exceptional. They were all accompanied by northeast winds, with surface temperatures at or a little above freezing, while similar or slightly higher temperatures prevailed on Mt. Washington. When the deposit of ice is in small amount, sufficing only to glaze the surface on which it is formed, it is called a "silver thaw."

288. Relation of rainfall and forests. The latter sections of this chapter will make it clear that the contrast between forested and barren regions depends

on the general winds and the form of the land areas, all of which are permanent physical features of the earth, as far as human history extends. There is, however, a very general impression that the presence of forests increases the rainfall; and that the destruction of forests may cause the rainfall to diminish so much as to reduce fertile regions to arid sterility. It is not to be doubted that the clearing of forests causes great fluctuations in the volume of streams, especially in hilly or mountainous regions. The streams overflow at times of heavy rain, when the undelayed surface water rushes down the slopes, washing the soil along with it, flooding and clogging the valleys with water and sand, and thus devastating both high and lowland; the streams almost disappear in droughts when the unsheltered ground is dried and springs weaken their flow; but it is quite another matter to affirm that the amount of rainfall is altered by the destruction of forests. There are few actual measurements that can be appealed to, and these do not give definite answer to the problem; the evidence that can be obtained does not clearly support the popular belief.

Popular opinion is also disposed to believe in an increase in the rainfall of our semi-arid western Plains by means of tree planting and agriculture; but no evidence in the form of actual records has been adduced to prove this very hazardous conclusion. The often-quoted account of a wholesale planting of trees in Egypt in the early part of this century and a consequent increase of rainfall is untrue; no such artificial climatic change has been produced in that country, and none need be expected in this.

289. Artificial rain. It is claimed by some that extensive conflagrations promote rainfall by exciting a convectional overturning in the atmosphere, which then grows to a rain storm; and by others, that violent concussions, such as the firing of heavy artillery on battlefields, causes rainfall even in times of drought. The advocates of these theories have at times tried to provoke rain artificially. While it might perhaps be possible to hasten the overturning of an almost unstable atmosphere by a vast conflagration, it would certainly be very expensive to attempt to alleviate the unfavorable conditions of a persistent drought or the long dry season of an arid region by this method, and we need not expect to witness its successful application. As for the efforts to produce rain by firing dynamite and other explosives in Texas in 1891, the official account of these experiments gives every indication that only a few pattering drops of rain were caused by the explosions, and these only when heavy clouds were floating overhead; the rains that followed the longest series of explosions were to all appearances ordinary summer thunder storms whose path happened to carry them over the places where the explosions were fired.

290. Rainbows. When falling rain is illuminated by the direct rays of the sun, a rainbow or arc of prismatic colors is seen on the rain, with its center opposite the sun, and a radius of 40–42½°; red is on the outside of the arc, and blue on the inside. An additional or secondary bow is sometimes formed outside of the first, and of fainter colors; its radius is 50–54°, and its colors are inverted from their order in the primary bow. Rainbows are produced by a complicated process of refraction of sunlight as it enters and passes out of the rain drops, internal reflection of the light within the drops, and interference of the rays after leaving the drops. The proper understanding of the problem requires a careful study of optics.

Rainbows cannot be seen from lowlands when the sun is high above the horizon; their arc increases in length as the sun comes nearer the horizon; and at sunrise or sunset they may form a full semicircle. Observers on mountain summits often see a rainbow of more than a semicircle when rain is falling through the air below them. In our latitudes, rainbows are most common in summer afternoons on the rear of receding thunder showers; because it is under these conditions that strong and nearly horizontal rays from the sun fall on a heavy curtain of rain. Another hour of occurrence might be more common near the equator, where thunder storms commonly move to the west; if occurring about sunset, the bow would precede the rain instead of following it. Cyclonic rains seldom produce rainbows; their rain area is generally followed by so large an area of cloud before the clear sky appears that the sun and the rain are seldom visible at the same time near the opposite horizon points; these larger rains have not the habit, so pronounced with our thunder storms, of moving away to the eastward about the time of sunset, and leaving a brilliantly clear sky immediately behind them.

291. Correlation of rainfall with the circulation of the atmosphere. It appears from the preceding explanations that rainfall is in practically all cases produced by some movement of the winds. We shall therefore now consider the distribution of rainfall over the world in quantity and in season of occurrence, in connection with the general and more local circulation of the atmosphere. We gain from this comparison a fuller appreciation of both processes than if either is considered alone; at the same time, it will be seen that the theory of the general winds, presented in Chapter VI, receives additional support from the explanation that it gives of the distribution of wet and dry regions and seasons. The rational correlations of the facts of temperature, pressure, winds and rainfall constitute the strength of the science.

292. Equatorial rains. The calms of the doldrums have already been described as occupying the belt of low pressure around the equator. They are supplied from either side by the inflowing trade winds which loiter here on

the weakest gradients, allowing the air to reach a high temperature and humidity and to expand upward, causing an overflow aloft, and thus establishing the upper currents that run obliquely towards the poles. In the process of expansion, there is also a diurnal convectional movement, caused on the equatorial lands by the warming of the lower air, but on the oceans presumably due in greater part to the upward expansion and diffusion of the plentiful vapor there taken from the water surface. In this belt of warm damp air, the noonday witnesses the production of clouds, followed in the afternoon or evening by the occurrence of lively showers of rain, which frequently reach the activity of violent thunder storms; late in the night the clouds dissolve away, and in the morning the sky is generally clear. The belt of doldrums is therefore also known as the equatorial cloud belt and as the belt of equatorial rains, standing in strong contrast with the comparatively dry trade wind belts on either side. The equatorial rainfall is estimated at about 100 inches. Its large amount is due not only to the activity of the convectional processes on which it depends, but also and largely to the rapid decrease of the capacity for vapor when air cools at the high temperatures prevailing around the equator.

293. Trade wind rains. The trade wind belts over the oceans, although of a rather high relative humidity, have a comparatively light rainfall because the temperature of their winds rises as they flow, and their capacity for vapor correspondingly increases. The evaporation that they cause from the ocean surface is so strong that a slightly greater degree of salinity is recognized in the ocean waters within their limits, the trade wind belts being separated by a belt of less saline water under the heavy fall of the equatorial rains. This is illustrated in Fig. 105, where the surface water of the trade wind areas has densities of more than 1.0270; while the equatorial belt is below 1.0265 or 1.0260.

If the trade wind encounters a mountainous island or a bold continental coast, the driven ascent of the air over such obstructions requires it to cool by expansion, thus producing clouds and generally rain as well; precipitation of this kind is known as tropical rainfall. For this reason, the windward slopes of lofty tropical islands, like those of the Antilles, and the windward coasts of torrid lands, like Guiana and southeast Brazil, are well watered, receiving a rainfall of from sixty to a hundred or more inches annually; while the leeward slopes are comparatively dry, as in Peru and northern Chile on the leeward slope of the Andes. The two sides of the Hawaiian islands are similarly contrasted, one being well clothed with tropical vegetation, and the other being comparatively dry and barren. In central America, the contrast between the well-watered and heavily-forested eastern slopes and the drier and more open western slopes has in great part determined the abandonment of

the former to the aboriginal tribes, while the latter have become the seat of European settlement.

An occasional cause of heavy rainfall in certain parts of the trade wind belts is found in the furious cyclones that traverse them on curved paths from the doldrums to the temperate zone. Although of not frequent occurrence, these cyclonic rains are truly torrential, as has been stated in Section 218.

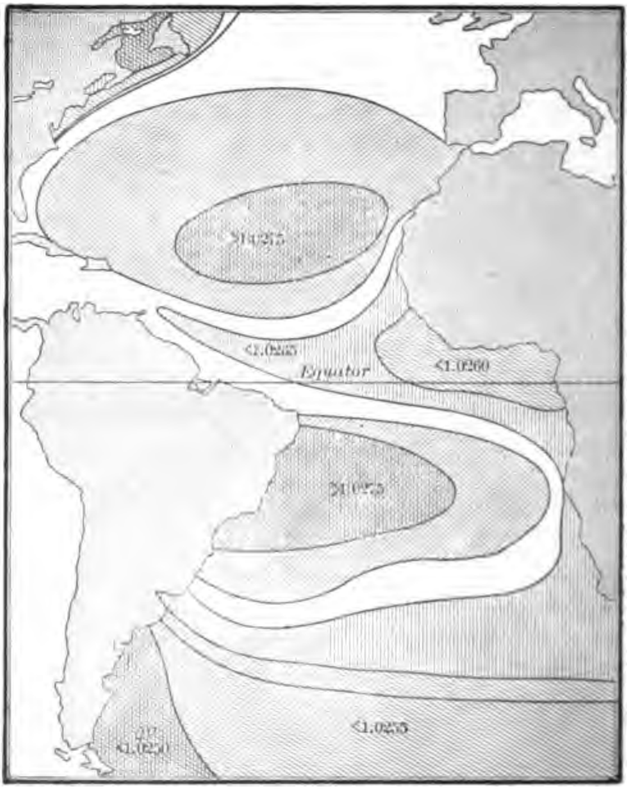

FIG. 105.

Data are not at hand to define closely the amount or distribution of rainfall from this cause; but on the Caribbean sea and in other cyclonic regions on the ocean, a considerable share of the total may be thus produced. Some of the Lesser Antilles have their greatest rainfall in the autumnal cyclone season.

294. Trade wind deserts. When the trade winds blow over a land of moderate elevation, they generally reduce its surface to a desert by depriving it of moisture; for it is not the quality of the desert rock or soil that makes

it barren, but simply the aridity. Plant life is almost or entirely driven away; the unsheltered dust produced by rock disintegration is carried off by the wind, and only sand, stones and rocky ledges remain. Thus the great Sahara, crossed by warming and drying winds from southern Europe and the Mediterranean towards the equator, has extremely little rainfall and is left a barren waste; excepting in the more lofty mountainous parts of its surface, where the ascending wind becomes rainy. Arabia, Persia, and a large part of Australia are sterile for similar reasons.

The greater area of deserts in the eastern than in the western hemisphere is the result of the outline of the lands and the trend of the great mountain chains. In the eastern hemisphere, the greatest breadth of Africa lies under the northeast trade winds; it is a desert plateau of moderate height with few mountains; a similar desert surface, but more broken by mountains, is continued within the trade wind belt across Arabia and Persia into northwest India. In the western hemisphere, the corresponding area north of the equator is largely oceanic, the American continent being narrowest in the latitudes of the northeast trades, and widest under the equatorial rains. Moreover, Mexico and Central America possess mountains and table lands of a considerable altitude, lying directly across the course of the winds, and thus calling forth a plentiful rainfall on the windward slopes at least.

South of the equator, Africa has a good supply of rainfall on its mountainous coast to the southeast, where the moist southeast trade from the Indian ocean strikes the land; but it contains a large area of moderate rainfall in the interior, and towards the western coast the desert of Kalahari repeats the aridity of the Sahara, but on a smaller scale. Australia presents a similar arrangement, having a narrow, well-watered coastal strip on the southeast, while the interior is for the most part too dry for occupation. A corresponding succession of parts may be seen in torrid South America, but with certain differences. South of the equator and towards the Atlantic coast, there are plentiful rains on the mountain slopes; further inland the country becomes lower and much drier; it is almost a desert in the trade wind latitudes near the eastern base of the Andes, but this great barrier again provokes rainfall and leaves only a narrow desert strip along the Pacific coast.

295. The horse latitudes or belts of tropical high pressure have been explained as regions of gently descending air, whence the trades and the surface members of the prevailing westerlies move away obliquely on either side. As ascending air cools and becomes cloudy and rainy, so descending air warms and becomes dry and clear. The horse latitudes are therefore regions of fresh clear air, drier than the trades and with little rainfall in their light and baffling breezes. The contrast between the equatorial and tropical belts of light winds and frequent calms is therefore highly instructive when con-

sidered in connection with the general circulation of the atmosphere; one being sultry, damp, cloudy and rainy; the other fresh, clear and relatively dry; just as the theory of Chapter VI would require. The growth of convectional clouds may produce local rains within this belt, but they are neither so plentiful nor so frequent as the heavy daily rains of the equatorial belt.

296. The stormy rainfall of the westerly winds. The westerly winds follow on the poleward side of the horse latitudes. These seldom produce rain from their own action, but they are subject in both hemispheres to frequent stormy or cyclonic overturnings, and in these overturnings the vapors gathered by the winds from the oceans are condensed to cloud sheets and yield plentiful rain. In certain parts of this belt, the precipitation is greater and more frequent in amount in winter than in summer, because in winter the activity of the winds is greater and the violence of the storms is then increased; this is especially apparent on the oceans and on western coasts in middle and higher latitudes. In other parts of this belt, particularly over continents at a distance from the oceans, where the continental indraft of the warm season draws damp air from the seas and where the high temperature then prevalent provokes local convectional storms, the rainfall is greater in summer than in winter. Thus the contrast between the greater winter rainfall of Oregon and Washington (state) in winter and the greater summer rainfall of the upper Mississippi valley is explained. A similar contrast is found between the rainfall of the western coast of Europe and of the interior plains of Russia and western Asia.

Rainfall is generally of sufficient amount in the belt of westerly winds over the oceans as well as over a great part of the lands, varying from thirty to eighty or more inches. Exception must however be made of continental interiors, distant from the oceans or enclosed by high mountains, where these middle latitudes are arid; and of bold western coasts of higher latitudes, where the rainfall becomes excessive, reaching more than a hundred inches at points where the form of the rising land gathers in the wind and locally increases the precipitation. Even so moderate a relief as that of Great Britain shows a decidedly greater rainfall on its western slopes, where the moist winds from the Atlantic first meet the highlands, than on the lower eastern slopes, where the winds flow after having lost some of their vapor. The Scandinavian peninsula shows the same contrast more distinctly; and it is exhibited with extreme emphasis in our western territories, of which more below. It must however be borne in mind that as the westerly winds often blow over mountains without yielding rainfall, while the cyclonic storms within these winds give forth rain not only on mountains but on lowlands also, the storms and not the action of the mountains on the general winds must be regarded as the controlling cause of precipitation in this belt; while

the mountains serve locally to increase the precipitation that the storms produce.

It is chiefly to cyclonic storms that the ample rainfall of the eastern United States is due. There is a rainfall of from 30 to 60 inches or more from the 96° or 98° meridian to the Atlantic coast, distributed with remarkable uniformity over this great region, especially in the growing season, and well apportioned through the year. Decidedly the greater part of this comes from cyclonic storms. The amount increases towards the Gulf of Mexico and the Atlantic coast, whence nearly all the supply of vapor for this rainfall is derived. In the Mississippi valley, there is a certain excess of the summer fall over that of the winter, mostly the product of local convectional storms, whose opportunity is found chiefly in a certain part of the cyclonic area; but this is on the whole an advantage to agriculture. Although droughts sometimes afflict considerable districts, and floods occasionally devastate the larger valleys, yet the world hardly contains as large an area as this so well adapted to civilized occupation. The importance of the warm waters of the Gulf of Mexico and of the western part of the North Atlantic eddy (including the Gulf Stream) cannot be overestimated in this respect. Instead of our having an American Sahara to the south of us in the trade wind latitudes, we have a great re-entrant of the oceanic shore line, into which flows a strong branch from the vast equatorial current of warm waters. The general winds, turned into an imperfect eddy around the North Atlantic basin (Sect. 157), here gather abundant vapor and shed it in a beneficent rainfall over the eastern half of our country. No formidable mountain range drains the winds of their moisture on their way inland, leaving the region to leeward a desert, as in southern Asia. Had North America been broad in the trade wind belt and narrow further north, its value as a home for man would have been greatly diminished.

297. Arid regions of the westerly winds. The southern part of South America is the only considerable land area in the belt of westerly winds in the southern hemisphere; its narrow western slope has abundant rains, while its broad eastern plains are comparatively dry; but being for the most part open to the adjacent Atlantic, they have a small or moderate rainfall from passing cyclonic storms. In the northern hemisphere, on the other hand, the continents expand to their greatest breadth in the latitude of the westerly winds, and include arid or desert regions of vast extent. The greatest of these extends over the western plains or steppes and the central basin of Asia. The steppes lie so far to the leeward of the Atlantic that the greater part of the vapor brought from that ocean has been condensed on the way, and the remainder is not easily prompted to fall. The interior basin, hemmed in on all sides by lofty mountains, is an extremely arid region; the rivers descending the interior

slopes from the snowy ranges weaken as they emerge on the piedmont plains and disappear further on in the sands of the desert.

In North America, the close approach of our Cordilleras to the Pacific, whence the westerly winds bring their vapor, leaves a large interior region with deficient rainfall. While the higher mountain crests and plateaus receive a relatively plentiful rainfall, bearing heavy forests above 7,000 or 8,000 feet altitude up to the tree line (about 10,000 feet), the plains between them are for the most part extremely dry and barren, and agriculture is limited to localities where irrigation from mountain streams can be introduced without too great expense. The driest part of this interior region lies in Arizona and in the part of southern California east of the higher mountains; here the rainfall averages less than three inches a year at several stations. Further eastward and northward, the rainfall gradually increases; but the influence of the Cordilleran rain shadow is felt half way across our continent.

298. Contrast of torrid and temperate rainfalls. A marked contrast is found between the distribution of rainfall under the easterly trades of the torrid zone and under the stormy westerly winds of the temperate zone. In the former, the occurrence of cyclonic storms is a minor feature, and rainfall is prompted chiefly by local storms or by the mountain ranges on the path of the winds. In the latter, cyclonic storms are the rule, especially in winter; mountain ranges are truly important in determining localities of greater and less rainfall, yet cyclonic disturbances are the chief rain makers. This is best seen in contrasting the prevalent fair weather of the trade belts in the broad Pacific with the inhospitable stormy weather of the high southern latitudes, where the westerly winds encircle the earth with hardly an interruption. The few expeditions that have penetrated that forlorn region bring reports of its continuous succession of blustering storms, with clouds and rain or snow; the unhappy product of an excessively large ocean surface, comparable in its depressing effects only with the arid interior of Asia, where the land area is excessive.

A further contrast is found between the rainfall of torrid and temperate latitudes in the occurrence of heavy rainfalls generally on the eastern coasts or mountain slopes of the former, and on the western coasts or slopes of the latter. Torrid western coasts are wet chiefly where they receive the equatorial rains, as in the Gulf of Guinea and on the Pacific slope of Colombia, South America; both of these regions lying somewhat north of the equator on account of the unsymmetrical position of the heat equator. In South America, the eastern coast in Guiana and Brazil and the eastern slope of the torrid Andes are well watered; while the western slope in Peru and northern Chile is dry. A similar arrangement is exhibited in Africa south of the equator, but to the north, the continuity of land towards Asia prevents its occurrence. On

the other hand, British Columbia and Alaska in the new world and the western coast of northern Europe in the old world are the regions of heaviest rainfall in the north temperate zone; while less rain falls to the eastward. Similarly, Patagonia has abundant rainfall on the side towards the Pacific, but is drier towards the Atlantic. Even in the Australian continent, the same relation appears; the Australian Alps having the most rain on the southeastern slope; while in Tasmania as well as in the islands of New Zealand the rainfall is received chiefly on the western slope.

An exception to the general arrangement of rainfall in the torrid zone, as stated above, is found in India and in the Malay peninsula. There the peculiar régime of the monsoons causes the western, not the eastern, coasts to receive the heaviest rain; and the double season of cyclones gives a larger share of cyclonic rainfall than is known elsewhere within the tropics, as is further explained below.

299. Rainfall of high latitudes. In passing further towards the poles, the proportion of snow in the total rainfall increases, but the total precipitation decreases, and in the polar regions it is comparatively moderate as far as observation goes. The annual sum is generally less than fifteen inches, and in certain polar regions it is less than ten inches. This is to be explained partly by the absence of local convectional storms, on which so much rain in the torrid zone and in the summer season of the temperate zones depends; and still more by the slow decrease of the capacity for vapor at the low temperatures of the polar regions; so that in spite of active cyclonic storms, the precipitation that they can yield is of small amount. A mass of saturated air cooled from 90° to 80°, as might happen in an equatorial thunder storm, would yield twenty times as much rainfall as if it were cooled from 0° to —10° in a polar cyclonic storm. Mention has already been made of the absence of snowflakes when the precipitation of the polar regions takes place at temperatures under 5° or 10° below zero; the precipitation then is in the form of fine ice spicules.

300. Migration of rain belts. The terrestrial wind system shifts north and south in an annual period following the passage of the sun. The belt of equatorial rains therefore moves north and south with the doldrums, as already illustrated by Fig. 59, for the torrid Atlantic, taken from the Pilot Charts of the North Atlantic, published by our Hydrographic Office at Washington, and from other sources.

At certain stations near the equator, the migration of the doldrums produces two rainy seasons with the sun overhead, and two intervening dry seasons when the sun is to the north or south. This is shown for several stations in the following table; São Thomé being an island in the Gulf of Guinea off Africa;

the Gaboon being a part of the equatorial African coast near by; and Quito and Bogotá being in equatorial South America.

301. Sub-equatorial rains. As a consequence of this migration, each trade wind belt is annually encroached upon by the equatorial rains as the sun enters its hemisphere. Thus the luxuriant forests of equatorial Africa merge into the wastes of the sandy Sahara through the more habitable belt of the Soudan, with dry winters and wet summers; the rainfall being less and its duration becoming briefer as the permanent desert area is approached, where practically no rain falls. South of the equator there is a corresponding arrangement of wet and dry seasons. A matter so important to ancient civilization as the flooding of the Nile depends on this control of the seasonal distribution of rainfall: when the sun is south, the Blue Nile rising in the mountains of Abyssinia is almost dry; but as the sun comes north and the calms and the rains follow it, the river rises and its volume is added to the more constant supply of the White Nile, which comes from the great lakes of equatorial Africa; thus causing the summer flood of the trunk river in middle and lower Egypt.

In South America, the extended plains or llanos of Venezuela are well watered from May to October when reached by the equatorial rains, but are dry and parched for the rest of the year, when they are swept over by the cloudless trade wind. While the llanos are dry, the interior campos of Brazil have their rains from October to April, and for the rest of the year they in turn have clear and dry weather. These may be called the regions of sub-equatorial rains.

The alternation of rainy and dry seasons is even more apparent in the monsoon region of India. When the sun is south, the northeast or winter monsoon flows from the mountains to the sea, and the greater part of the vast peninsula is dry. In the opposite season, when the high temperatures of early summer have shifted the belt of low pressure to northern India and Persia, the southeast trade wind comes across the equator and becomes the southwest or summer monsoon, causing heavy rains on the Ghâts or bold western coast of southern India, on the mountains of Burmah, and on the lofty Himalaya further north. On the latter, the greatest rainfall of the world is found, amounting to forty feet a year north of the Bay of Bengal, and nearly all of this falls in the five months from May to September (see Sect. 284). The southeastern coast of Asia has for the same reason a contrast between wet summers and drier winters, but the difference is less marked than in India. Northern Australia, with a northwest landward monsoon in the southern summer, has its rainy season from November to March or April.

The following table gives the precipitation of certain torrid stations in inches to illustrate the amount and distribution of equatorial and tropical rainfall. The pages in the first column refer to Hann's *Klimatologie*.

HANN.	LOCALITY.	LAT.	Dec.	Jan.	Feb.	Mar.	April.	May.	June.	July.	Aug.	Sept.	Oct.	Nov.	YEAR.
	Central and S. America.														
348	Guatemala	15° N	0.4	0.3	0.1	0.8	3.0	5.6	11.1	10.8	8.9	8.9	7.2	0.5	57.5
"	Panama	9° N	2.8	0.5	0.8	1.6	2.8	7.6	7.9	7.6	6.8	7.5	9.5	11.6	66.9
350	Bogotá	5° N	7.5	5.4	4.2	3.5	9.5	7.4	3.9	3.6	4.6	3.4	9.1	11.8	73.9
"	Quito	0° –	2.8	3.2	6.4	4.4	6.6	5.1	2.5	1.4	1.8	1.8	4.4	6.2	46.6
349	Havana	23° N	2.2	3.3	1.6	1.5	3.2	4.1	5.7	4.9	4.8	6.0	6.7	2.2	46.3
348	Caracas	10° N	1.0	0.4	0.3	0.3	1.0	2.2	4.6	4.8	3.6	5.4	5.0	2.6	31.1
350	Georgetown	7° N	10.7	6.8	5.8	7.3	7.3	14.1	13.1	8.2	7.6	2.9	3.3	4.1	93.5
350	Rio Janeiro	23° S	5.2	5.4	4.7	5.9	3.4	4.8	1.5	1.3	2.8	3.3	3.9	5.7	47.8
	Africa, west coast.														
253	St. Louis	16° N	0.0	0.2	0.5	0.0	0.0	0.2	0.4	2.6	8.0	3.7	0.5	0.0	16.2
253	São Thomé	0° –	3.1	4.2	4.8	7.2	4.6	5.1	0.9	0.0	0.6	0.7	5.0	5.9	42.0
"	Gaboon	0° –	7.5	7.5	9.0	15.0	8.7	8.9	0.9	0.5	1.2	8.1	19.4	19.1	105.8
"	Loanda	9° S	1.2	2.4	1.1	1.3	3.3	0.3	0.0	0.0	0.0	0.1	0.2	2.5	12.5
	Africa, east coast.														
253	Mombas	4° S	1.9	1.6	1.7	3.4	7.8	12.3	4.9	5.2	3.6	3.0	4.9	5.6	55.8
279	Port Louis	2° S	3.7	5.7	11.8	5.2	3.1	2.1	1.5	0.9	1.5	0.4	0.7	1.7	38.3
	India.														
295	Colombo, Ceylon	7° N	5.8	3.3	1.6	5.8	9.6	13.9	8.4	5.7	4.2	5.2	13.4	11.2	88.3
296	Bombay	19° N	0.0	0.1	0.1	0.0	0.0	0.6	20.9	24.3	15.4	10.6	1.6	0.5	74.1
	Mahableshwar	18° N	0.3	0.4	0.1	0.4	0.9	1.5	46.2	95.7	71.6	3.2	5.0	1.0	254.3
	Delhi	29° N	0.4	0.8	0.6	0.8	0.4	0.7	2.8	8.4	7.2	4.4	0.7	0.1	27.6
	Cherrapunji	25° N	0.2	0.8	2.8	8.8	30.9	51.4	115.9	130.8	79.6	56.1	13.7	2.2	493.1
	Southeastern Asia, etc.														
297	Hong Kong	22° N	0.6	0.4	1.5	2.6	3.7	9.5	17.2	14.2	12.5	13.9	5.7	2.9	84.6
324	Singapore	1° N	10.8	8.4	6.1	6.8	7.0	6.3	7.2	6.2	9.1	7.1	8.8	10.7	94.5
"	Port Darwin, Australia . .	12° S	11.4	14.9	11.5	12.8	4.7	0.8	0.0	0.0	0.0	0.0	0.6	0.9	40.6

weather in summer, after the fashion of the torrid zone; yet even in this season cyclones have a considerable effect. In winter, the diurnal control weakens, especially when the ground is snow covered, and the cyclonic control strengthens, so as to determine nearly all the changes from clear to cloudy, from warmer to colder, from calm to windy, from wet to dry. Examples of the weather in each season may now be given.

314. Summer weather in the central United States. The warm spells of summer time occur during the gradual advance of a moderate cyclonic area over the upper Mississippi valley. A light southerly wind — a warm wave, or sirocco — prevails on moderate gradients in front of the center of low pressure. There is at first a strong diurnal range of temperature, with a quick warming in the morning (see Fig. 10*a*) and five or six hours of high temperature during the later half of the day. As the sky becomes more hazy or more streaked with cirrus clouds, the maximum temperature reaches a higher and higher degree each day, and the nocturnal cooling diminishes; the air becomes sultry and oppressive with increasing humidity; the ground is parched and the wind drifts clouds of dust from all bare surfaces; vegetation is stifled; men and beasts suffer greatly while at labor, and sunstrokes are reported in increasing numbers from the crowded cities. Unless a rain has recently fallen, the sky may be nearly cloudless all this time; but near the culmination of the warm wave, cumulus clouds may grow to the size of local thunder storms in the afternoon, trailing a refreshing rain beneath them; yet the temperature rises again after their passage. Near the center of the cyclonic area, there are clouds and rain; further south along the trough of low pressure, there are extended thunder storms; and after these pass by, a welcome shift turns the wind to the west and northwest; the temperature falls ten or twenty degrees, the air becomes fresh and pleasant, and the sky brightens to a clearer, darker blue. If the rainfall by which the hot spell was terminated comes in the afternoon or at night, as is often the case, there is active evaporation the next morning under the drying northwest wind — the cool wave of summer — and a slow rise of temperature to a moderate maximum late in the afternoon; the morning sky is early flecked with growing cumuli, and by noon it may be overcast above a brisk wind; but the night will be clear and calm again, and the next day will be less cloudy as the ground dries. Then the temperature increases day by day; and generally by the third day or sooner, the winds weaken on the faint gradients of the anticyclonic area which then passes by; the sky still being fair and the range of temperature strong, giving warm days and pleasant nights. As soon as the pressure begins to fall on the western side of the anticyclone, the wind swings around to southerly again, and another warm spell sets in. During the whole of such a period as this, the diurnal changes are perfectly apparent, and for a part of the time they are

Station.	Latitude.	Dec.	Jan.	Feb.	Mar.	Apr.	May.	June.	July.	Aug.	Sept.	Oct.	Nov.	Year.
San Diego, Cala.	33° N.	2.34	1.96	2.36	1.47	0.87	0.36	0.06	0.01	0.15	0.07	0.47	0.82	10.98
San Francisco, Cala. . .	38° N.	5.11	4.98	3.72	3.28	2.14	0.68	0.27	0.02	0.01	0.20	1.34	2.79	24.30
Neah Bay, Wash. . . .	48° N.	14.45	17.29	11.14	8.08	5.12	4.49	5.00	2.72	2.28	6.59	11.52	12.92	101.87
Sitka, Alaska	57° N.	10.11	9.75	10.51	10.02	6.24	4.94	3.58	5.28	6.93	11.09	13.49	13.68	111.72
Salt Lake City, Utah . .	41° N.	1.59	1.52	1.38	1.92	2.36	1.78	0.75	0.51	0.81	0.83	1.70	1.43	14.52
Omaha, Nebr.	41° N.	0.96	0.68	0.76	1.47	3.08	4.52	5.87	5.32	3.39	3.21	2.60	1.22	32.94
New Orleans, La.	30° N.	4.64	5.40	4.30	5.68	5.38	5.32	6.77	6.42	6.24	4.89	3.42	4.46	62.94
New York, N.Y.	41° N.	3.33	3.89	3.76	4.10	3.40	3.05	3.41	4.57	4.75	3.94	3.46	3.69	45.47
Jacksonville, Fla.	30° N.	3.03	3.45	3.16	3.26	3.12	4.41	5.61	6.60	6.60	7.93	5.68	2.61	55.95
Funchal, Madeira Islands .	33° N.	4.4	6.4	2.9	2.7	1.5	1.1	0.6	0.0	0.3	1.1	2.7	5.5	29.2
Coast of Algeria	35° N.	5.0	3.3	3.0	3.9	2.2	1.4	0.8	0.0	0.3	1.1	3.0	3.6	27.6
Alexandria, Egypt . . .	31° N.	2.3	2.2	1.5	0.9	0.1	0.0	0.0	0.0	0.0	0.0	0.4	1.2	8.6
Western England[1] . . .	53° N.	3.6	4.0	2.9	2.4	2.3	2.2	2.5	2.9	3.1	4.1	4.0	3.2	37.1
Middle Germany	50° N.	2.2	1.0	1.4	1.6	2.1	1.4	2.4	2.6	2.2	1.5	1.5	2.3	22.2
Southeast Russia	50° N.	0.9	0.9	0.8	0.8	1.1	1.7	2.3	1.7	1.5	1.4	1.1	1.2	15.4
Western Siberia	50° N.	0.7	0.5	0.5	0.5	0.7	1.0	2.5	2.6	2.3	1.3	1.0	1.0	14.6
Cape of Good Hope . . .	36° S.	0.9	0.6	0.9	0.9	1.8	4.6	5.8	4.7	4.1	2.8	2.1	1.5	30.7
South Coast, Australia . .	37° S.	1.4	1.1	1.1	1.4	1.9	3.2	3.5	3.5	3.8	3.0	2.2	1.6	27.2
Central Chile	34° S.	0.2	0.0	0.2	0.5	0.5	2.6	3.8	3.4	2.2	1.4	0.6	0.3	15.7
Southern Chile.	41° S.	4.5	4.5	3.4	7.9	10.2	14.8	18.2	18.2	13.6	6.8	5.7	5.7	113.5
Interior, Argent. Repub. .	32° S.	3.3	3.9	3.1	2.7	1.6	0.2	0.4	0.2	0.4	0.6	1.4	2.7	20.5
Buenos Aires	37° S.	2.6	2.8	2.6	2.3	2.1	1.8	2.1	1.6	1.0	1.8	2.6	2.6	25.9

[1] The monthly values for districts, in this and the following examples, are not to be taken as exactly equivalent to the means at any single station, but only as representing the average part of the total for the district that falls in a single month.

Lower California, where very little rain falls at any time. We find in this the full explanation of the seasonal distribution and the increase of rainfall northward along the Pacific coast; from 12 inches at San Diego, to 24 at San Francisco, 83 at the mouth of the Columbia river and 105 at Neah Bay, Washington.

A similar arrangement of seasonal rainfall is found in Chile. The southern extremity of that country, commonly known as the western slope of Patagonia, is rainy throughout the year. Northern Chile is persistently dry, forming the desert of Atacama. An intermediate portion is well watered in winter, when the stormy overturnings of the westerly winds reach it, and dry in summer, when it is brushed over by the warming branch winds from the westerlies to the trades.

Northern India is peculiarly situated with respect to the sub-tropical rains. In the winter season, rains from weak cyclonic storms occur on the northern plains and along the marginal ranges of the Himalaya, progressing from west to east, after the usual fashion of the cyclonic storms of the temperate zone, and thus indicating their correspondence with the winter rains of other sub-tropical regions. But the opposite or summer season, instead of being dry, is wetter than the winter season, because at this time rain falls from the storms of the summer monsoon.

A review of the preceding paragraphs may be made in the table on the preceding page, in which several characteristic rainfalls of the sub-tropical belt and of the westerly winds are assembled. The examples for the United States are supplied by the national Weather Bureau; most of the rest are selected from Hann's *Klimatologie*, in which many other records may be found.

303. Effect of clouds and rainfall on the general circulation of the atmosphere. Section 199 has explained the greater altitude reached by cloudy than by clear local convectional currents. An extension of the same principle will explain the assistance given to the general circulation of the atmosphere by the liberation of latent heat in the rainy cloud belt around the equator. The retarded cooling of the ascending currents in which the clouds are formed enables the air at a given altitude to maintain a higher temperature than it would possess if clear; and as the equatorial clouds are chiefly formed in the day-time, their presence arrests a certain share of insolation high above sea level. The relative increase of temperature thus determined causes the isobaric surfaces to diverge more strongly towards the equator than they otherwise would, and thus hastens the general circulation. The absence of high temperatures in the lower air over the torrid seas is in this way in part made good, as far as the terrestrial winds are concerned.

The abundant cloudiness and frequent rainfall of high latitudes does not counter-balance the effect produced by the rain clouds around the equator; for

at the low temperatures prevalent far north and south the latent heat liberated is not great, and the cooling of the air is but slightly retarded. Furthermore, as terrestrial radiation is relatively strong in high latitudes, the presence of clouds there must allow the air to cool more than if it were clear, and hence the isobaric surfaces will converge poleward more rapidly than they would otherwise. It therefore seems, on the whole, that cloud-making and rainfall must accelerate the terrestrial circulation.

The co-operation of sea and valley breezes has been already mentioned. It may now be perceived that the clouds formed and the rain falling in the day-time over mountainous islands must still further aid the flow of the diurnal winds. In the same way, the heavy monsoon clouds and rains of India in summer must help along the inflowing monsoon winds. The circulation around the low pressure areas of the northern Atlantic and Pacific oceans would, on the other hand, appear to be retarded in winter by the radiation from their extensive cloud masses, whose action in this respect must tend to diminish the excess of temperature that would otherwise prevail on account of the abnormal warmth of the ocean waters and the liberation of latent heat during the formation of the abundant clouds of these regions.

CHAPTER XIII.

WEATHER.

304. Weather. Thus far, our attention has been directed to the study of phenomena in their complete and ideal development; the general distribution of temperature over the world; the terrestrial and continental circulations of the atmosphere; the larger and smaller storms; all these have been described according to their conditions of occurrence, following each one over all the area that it occupies. Similar phenomena in different parts of the world have been considered together; the low pressure at the south pole with that at the north pole; the trade wind belts all around the torrid zone; the continental winds of Australia with those of North America; the cyclones of the various tropical oceans; and so on.

In this chapter, we proceed in an entirely different order, and consider the actual succession of phenomena, however varied and arbitrary, as they are experienced at one place or another in different parts of the world. Some suggestion of special effects of this kind has been presented in the accounts of the passage of a tropical cyclone (Sect. 218), of the weather changes caused by the passage of cyclones in our latitudes (Sect. 243), and of the passage of thunder storms (Sect. 252); but these sections had for their first intention the better understanding of the phenomena that they considered, rather than the part that these phenomena play in determining the succession of events that an observer at any single station would experience. We therefore now turn more particularly to the study of events in the natural order in which they happen at any single place; the successive meteorological conditions of all kinds being collectively named by the term, Weather.

305. Weather elements. All the atmospheric conditions which an observer notices by sight or feeling may be called weather elements. These include the temperature, humidity and movement of the air about him, the condition of the sky as to clouds or haze, and the occurrence of precipitation, as rain, snow or hail. Inasmuch as observations are commonly carried on near sea level, it follows that weather is largely concerned with the conditions of the lower air; but the weather noted by an observer on a high mountain peak would be governed by the lofty currents. Besides the familiar and sensible elements above named, it is convenient to include another, of which our unaided senses give us no indication. This is the pressure of the atmosphere, as determined by the barometer. Although not properly a weather element itself, it is of so much importance in understanding the relations of the actual

weather elements and the meaning of their changes, that it is advisably considered with them.

306. Control of weather changes. The change of weather from one condition to another may be recorded by its effects on the various instruments already described for determining temperature, wind, and so on. This may be done without reference to causes and without any search for explanation of change; but it is not the object of this book to encourage the keeping of so insufficient a record as such a one would be. Along with the careful, unprejudiced record of the facts of observation, there should always go a serious attempt to refer the facts to their causes, whether simple or complex. It is thus possible to gain an understanding of many changes, and even to anticipate their occurrence; but much remains to be discovered in this direction.

The chief controls of weather changes are, first, the diurnal variation of insolation from day to night; second, the annual change of insolation from summer to winter; third, the passage of cyclonic and anticyclonic areas, with their attendant smaller storms; fourth, the passage of irregular surges of pressure and temperature, which appear to be of longer period, larger area and slower eastward movement than cyclones and anticyclones are. The diurnal and annnal controls are regular in period and comparatively regular in value at any single station; they vary in intensity from place to place in accordance with such factors as latitude, form and altitude of land, and distance from the sea; they are relatively weak on the torrid oceans, and strong on the inner lands of high latitudes. The other controls are everywhere more or less irregular in period, and they vary greatly in frequency and intensity in different parts of the globe.

The diurnal change gives us warm days and cool nights. The annual change brings us fair, warm summer weather and stormy, cold winter weather. Cyclonic and anticyclonic changes interrupt the regular sequence of diurnal changes and introduce the chief irregular alternations from cloudy, wet and windy weather to fair weather with relatively dry and quiet air. It is for the most part with these changes that weather prediction is concerned. Surges seem to control spells of weather that last one or several weeks, as in spells of unusual heat or cold at different times of year. Successive surges sometimes follow tolerably regular periods; these are easily recognized after their completion; they continue as long as they are not broken by some irregular variation of unknown cause; but no one has yet succeeded in determining beforehand how many surge periods may occur or when they will be broken into by some other cause of change: hence, for the present, the surge is not practically utilized in forecasting the weather. The combination of all controls produces an exceedingly irregular sequence of weather changes, which in our latitudes cannot at present be foretold for more than a few days in advance at most.

307. Weather of the torrid zone. The torrid zone, embracing about a half of the earth's surface, and including a large oceanic area, is for the greater part characterized by a regular sequence of distinct diurnal changes, with a comparatively steady and regular change of weather in passing from one season to the next; and further by small interference from cyclonic interruptions. One day is much like another; even the small double diurnal oscillation of atmospheric pressure is repeated day after day with trifling irregularity. The diurnal range of temperature is remarkably steady. The winds over the trade wind belt are much steadier in direction and velocity than we know them here. Clouds increase in the day-time and decrease at night. When rain occurs, as in the equatorial belt, it manifests a distinct diurnal period, falling chiefly late in the day, while the mornings are generally dry.

A closer study of the weather of this vast zone suggests its division, first, according to the general wind system; second, according to ocean and land areas. We thus recognize, first, the two belts of the steady trade winds; second, the equatorial belt, where the weather in different times of year varies with the change from the wet to the dry season. Each of these divisions should then be divided into continental and oceanic areas, as the intensity of weather changes from day to day and from season to season will be found much more pronounced on land than at sea.

308. The trade wind belts. In the trade wind belts at sea, the constancy of the winds and the regular succession of the moderate diurnal range of temperature, day after day, is hardly to be imagined by those who know only our stormy portion of the temperate zone. The wind flows from the same quarter and with about the same velocity day and night; clouds form in the day-time in moderate amount, and generally die away in the evening, seldom yielding rain, except near the doldrums, or when a cyclone passes by and takes the control of all weather changes into its own hands. The cooler weather of the year has a temperature only eight or ten degrees lower than the weather in the warmer months.

In the trade belt on land, the dryness of the air and the increase in the diurnal range of temperature and of wind velocity are the most notable features. In the hot season, the heat at mid-day is extreme, and when accompanied by high wind, bearing clouds of dust and sand, it may be fatal to human life. Yet the nights even in the hot season are calm and relatively cool. Travellers in trade wind deserts make frequent mention of the discomfort or suffering in the parching winds of day-time and the comparative comfort of the nights. In the colder season, the temperature at night falls to a much lower degree than we associate with the torrid zone; close to the ground, it may approach the freezing point, and thin sheets of water in shallow vessels, cooling by radiation and evaporation, may be frozen over. The equatorial margin of a

trade wind belt partakes of the features of the sub-equatorial belts, when the march of the sun brings the rains upon it. Here the weather at different seasons is strongly contrasted; varying from the extremes of warm, dry winds at one season, to sultry calms, broken by violent thunder storms and drenching rains at another. In the monsoon region of India, particularly in the northern plains of the peninsula, the contrast between the weather of different times of year reaches an extreme for the torrid zone. The weather in the cold season is cool and relatively dry, occasionally including distinct cyclonic changes, when an extra-tropical cyclone, moving eastward, causes non-diurnal variations of wind, temperature, cloudiness and rain. The early summer brings days that are hot enough for an equatorial latitude; and when the wet season succeeds the hot season, the dry heat is followed by weather as sultry and oppressive as that of the doldrums.

309. The equatorial belt. Between the trade wind belts, the weather of the doldrums is hot, moist, cloudy and sultry, with calms or light breezes and frequent rains, returning day after day with much regularity. When the doldrums move away and the trade winds follow them, the weather for a time becomes clearer and fresher. The equatorial belt on land is more pronounced in its weather types than at sea. Its heat is more intense; the storms of its daily rains appear to be more severe than at sea; its variation of weather from one season to another is much more marked, owing to the increased migration of the wind belts; but it is still essentially diurnal weather. An account of Java, by Junghuhn, calls especial attention to the surprising regularity of the daily weather changes in the interior of the island. The evening sky soon becomes clear, and the air relatively cool and damp. Late at night, fogs form on the lowlands. The morning sun dispels the fogs, but cumulus clouds soon form, particularly over the mountains, and a light breeze springs up. Rain often falls on the mountains in the afternoon, but soon after sunset, the clouds dissolve away again, and the series of changes is begun once more. All of this is purely diurnal weather in its perfection.

310. The sea breeze. On the coasts of equatorial lands, the distinct diurnal control of the weather causes the sea breeze to attain an importance that it never reaches with us. The old navigator, Dampier, described this with much appreciation two hundred years ago: —

"Sea breezes do commonly rise in the morning about nine a clock, sometimes sooner, sometimes later; they first approach the shore, so gently, as if they were afraid to come near it, and ofttimes they make some faint breathings, and as if not willing to offend, they make a halt, and seem ready to retire. I have waited many a time both ashore to receive the pleasure, and at sea to take the benefit of it. It comes in a fine small black curle upon the water,

when as all the sea between it and the shore not yet reached by it, is as smooth and even as glass in comparison; in half an hour's time after it has reached the shore it fans pretty briskly, and so increaseth gradually till 12 a clock, then it is commonly strongest, and lasts so till 2 or 3 a very brisk gale; about 12 at noon it also veers off to sea two or three points, or more in very fair weather. After 3 a clock it begins to dye away again, and gradually withdraws its force till all is spent, and about 5 a clock, sooner or later, according as the weather is, it is lull'd asleep, and comes no more till the next morning. These winds are as constantly expected as the day in their proper latitudes, and seldom fail but in the wet season."

"Land-breezes are as remarkable as any winds that I have yet treated of; they are quite contrary to the sea-breezes; for those blow right from the shore, but the sea-breeze right in upon the shore; and as the sea-breezes do blow in the day and rest in the night; so on the contrary, these do blow in the night and rest in the day, and so they do alternately succeed each other. For when the sea-breezes have performed their offices of the day by breathing on their respective coasts, they in the evening do either withdraw from the coast, or lye down to rest; then the land-winds whose office is to breathe in the night, moved by the same order of divine impulse, do rouze out of their private recesses and gently fan the air till the next morning; and then their task ends and they leave the stage."

"These land-winds are very cold, and though the sea-breezes are always much stronger, yet these are colder by far. The sea-breezes indeed are very comfortable and refreshing; for the hottest time in all the day is about 9, 10 or 11 a clock in the morning, in the interval between both breezes, for then it is commonly calm, and then people pant for breath, especially if it is late before the sea-breez comes, but afterwards the breez allays the heat. However, in the evening again after the sea-breez is spent, it is very hot till the land-wind springs up, which is sometimes not till 12 a clock or after."

These several paragraphs suffice to show that over large areas of the torrid zone the sequence of weather changes is so simple and so regular, and one day is so much like its neighbors, that almost any day is a fair sample of its season; and hence the average of the successive values of the weather elements, or climate, hardly differs from the observed values of individual occurrences, or weather. This will be even more apparent in the next chapter, when it is seen how much of what is here said might there be pertinently repeated. It is on the other hand to be expected that when closer attention is given to the individual weather features of the torrid zone, they may be found to be more complicated than they are here described.

311. Weather of the temperate zones. The middle latitudes of the earth are characterized chiefly by an irregular combination of periodic or diurnal,

and unperiodic or cyclonic and anticyclonic weather changes. The quality of weather produced by this combination varies greatly in different seasons of the year; the periodic changes predominating, especially on land, during the higher temperatures of summer and in the clear air of anticyclones; and the irregular changes being in control during the lower temperatures of winter. The quality of weather varies also very greatly according to the situation of the observer in the zone. Out of the abundance of examples that may be found, space can be given only to a few of the more peculiar ones: the oceanic areas, especially of the southern temperate zone; the land interiors and the western and eastern coasts of the northern temperate zone.

312. The south temperate zone. The great temperate water zone of the southern hemisphere is almost as regular in its constant succession of unperiodic weather changes as the trade wind belts are in their unchanging succession of constant fair-weather days. East-bound vessels on the southern seas sail day after day before boisterous winds, whose strength often rises to a gale, generally from some westerly quarter, shifting between northerly and southerly points, but seldom blowing from the east. The temperature is inclement; its average diurnal changes are small, for on the ocean day and night are light and dark, rather than warm and cold; but its irregular changes are pronounced though not severe; they follow the shifts of the wind rather than the rising and setting of the sun. Cloudy skies are prevalent and rain or snow often falls, but not in excessive amounts. In winter, the weather changes are stronger, more frequent and more distinctly cyclonic; they differ more in this way from the weather changes of summer than in a strongly decreased temperature. The accounts already given of cyclonic winds explain that in the southern temperate zone, the northerly winds bring milder temperature, clouds and rain; while the southerly winds follow with cooler and clearing weather. Little is known of surges here, although it is probable that they occur. As in the trade wind belt of the torrid zone, so in the southern temperate ocean belt, a brief period of observation will furnish a fair sample of weather for the season; but here the weather period is roughly two or three days, being the time required for the passage of a cyclone and an anticyclone, instead of the passage of a day and a night.

313. The north temperate zone. The weather over the broad oceanic areas of the north temperate zone is much like that of the corresponding southern zone in varying chiefly with the procession of cyclones and anticyclones by which its wind, temperature and sky are controlled; although in summer the strength of these controls is much weakened.

The great continental interiors of the north temperate zone, such as our vast Mississippi basin, are characterized by frequent spells of regular diurnal

weather in summer, after the fashion of the torrid zone; yet even in this season cyclones have a considerable effect. In winter, the diurnal control weakens, especially when the ground is snow covered, and the cyclonic control strengthens, so as to determine nearly all the changes from clear to cloudy, from warmer to colder, from calm to windy, from wet to dry. Examples of the weather in each season may now be given.

314. Summer weather in the central United States. The warm spells of summer time occur during the gradual advance of a moderate cyclonic area over the upper Mississippi valley. A light southerly wind — a warm wave, or sirocco — prevails on moderate gradients in front of the center of low pressure. There is at first a strong diurnal range of temperature, with a quick warming in the morning (see Fig. 10*a*) and five or six hours of high temperature during the later half of the day. As the sky becomes more hazy or more streaked with cirrus clouds, the maximum temperature reaches a higher and higher degree each day, and the nocturnal cooling diminishes; the air becomes sultry and oppressive with increasing humidity; the ground is parched and the wind drifts clouds of dust from all bare surfaces; vegetation is stifled; men and beasts suffer greatly while at labor, and sunstrokes are reported in increasing numbers from the crowded cities. Unless a rain has recently fallen, the sky may be nearly cloudless all this time; but near the culmination of the warm wave, cumulus clouds may grow to the size of local thunder storms in the afternoon, trailing a refreshing rain beneath them; yet the temperature rises again after their passage. Near the center of the cyclonic area, there are clouds and rain; further south along the trough of low pressure, there are extended thunder storms; and after these pass by, a welcome shift turns the wind to the west and northwest; the temperature falls ten or twenty degrees, the air becomes fresh and pleasant, and the sky brightens to a clearer, darker blue. If the rainfall by which the hot spell was terminated comes in the afternoon or at night, as is often the case, there is active evaporation the next morning under the drying northwest wind — the cool wave of summer — and a slow rise of temperature to a moderate maximum late in the afternoon; the morning sky is early flecked with growing cumuli, and by noon it may be overcast above a brisk wind; but the night will be clear and calm again, and the next day will be less cloudy as the ground dries. Then the temperature increases day by day; and generally by the third day or sooner, the winds weaken on the faint gradients of the anticyclonic area which then passes by; the sky still being fair and the range of temperature strong, giving warm days and pleasant nights. As soon as the pressure begins to fall on the western side of the anticyclone, the wind swings around to southerly again, and another warm spell sets in. During the whole of such a period as this, the diurnal changes are perfectly apparent, and for a part of the time they are

dominant; but they fall to a moderate value about the time of highest temperature, when even the nights are oppressively warm. The character of each succeeding day manifestly depends largely on its relation to the controlling cyclonic or anticyclonic center, and the consequent behavior of the cyclonic or anticyclonic winds. There are of course many occasions when the areas of low and high pressure are poorly defined and their effects are weak; and there are many days, especially in summer, when the distribution of pressure is indefinite, and it is not possible to locate any cyclonic or anticyclonic centers within the eastern part of our country; but by far the greater number of days may be characterized simply and accurately enough by referring them to one or another part of the passing areas of low or high pressure.

315. Winter weather in the central United States. In winter, the cold wave is the emphatic phenomenon. Beginning our record with a spell of clear, calm, cold weather, the ground being covered with snow, let us notice when the wind turns to a southerly source, springing up perhaps at night, and thus checking the nocturnal fall of temperature. The next day the temperature rises rapidly and a thaw sets in under warm sunshine; but as the day passes the sky clouds over, and by afternoon the sun is lost behind a dull bank of matted cirro-stratus. Yet the temperature continues to rise even into the night; the air is unseasonably warm; our heavy clothes and over-heated houses are oppressive; the thaw progresses and is aided by a fall of wet snow which soon turns to rain. The next morning is still rainy and dark under low clouds with misty air and a foggy ground; gradually the wind shifts to westerly; the rain ceases, the clouds break in the west or northwest, their edge drifts obliquely eastward, revealing a pure blue sky, the wind strengthens, and then even under bright sunshine the temperature begins to fall; the freezing point is soon passed, the half-melted snow is frozen again into an icy sheet; the night is cloudless, windy and bitter cold; the cold continues through the next day with hardly any rise of temperature under strong sunshine. Then the wind falls away at sunset, as the anticyclonic area comes on, and the next morning our neighbors report the most extreme cold in the valleys and lowlands, and more moderate temperatures on the hilltops — an anticyclonic inversion of temperature often amounting to twenty or thirty degrees.

An extraordinary instance of weather changes of the kind just described was recorded in New England on Christmas Eve, 1886. The sky had been cloudy under an uncomfortable southerly wind that was flowing toward a cyclonic center on its way down the St. Lawrence. The temperature had risen almost continuously from the evening of the 23d. During the evening of the 24th there was a heavy rainfall, and about midnight the highest temperature of the month (55°) was registered; a temperature that should have been felt, according to diurnal and annual controls, a little after noon on the first day of

the month. Shortly after midnight, the wind quickly shifted to the west; the temperature immediately began to fall, and continued to fall almost steadily all the next day under a clearing sky, thus ushering in a strong cold wave.

Changes of this kind, more or less pronounced, are so rapid in their succession that we seldom have more than a day or two of uniform weather in winter. However fair the morning, the sky may be half overcast with cirrus streamers by noon; and after sunset rain or snow may be falling. However low the clouds at one hour, twelve hours later may see them all dispelled. The variety in the succession of weather elements is endless; yet all the variations are on one theme.

316. Weather of the frigid zones. During much of the year in high latitudes, the diurnal control of weather is wanting. Near the equinoxes, when the sun rises and sinks a few degrees above and below the horizon, there is a semblance of the changes which we know so well; but at other seasons of continued night or uninterrupted day, the weather varies only through stormy and fair spells, presumably the effect of passing cyclones and anticyclones, as with us in the temperate zone. During the continued daylight, which Arctic travellers find extremely tiresome, the periods of weather change are no better defined than the hours of work and rest. Fair weather continues fair and bright until broken by a storm; stormy weather continues dull, snowy or wet until followed by clearing skies. In the long winter night, storms of wind and snow are followed by spells of quiet air and more extreme cold; but without more regular sequence than results from the irregular periods of cyclonic passage.

WEATHER OBSERVATION AND PREDICTION.

317. Weather observations. At the more important stations of the national weather services, at some of the larger astronomical observatories, and at certain private observatories, all the weather elements are carefully observed, frequently with self-recording instruments by means of which a full account of all changes is taken. At less important stations of national weather services, at stations maintained by volunteer observers of state weather services, or by independent private observers, records of temperature, wind, sky and precipitation are taken more or less completely once, twice, or three times a day. In all these cases it generally happens that the method of reduction of observations by averaging them in diurnal or monthly periods obliterates to a greater or less degree the irregular changes of the weather, such as result from the passage of cyclones and anticyclones, and gives undue emphasis to the effects of diurnal and annual controls. The record and reduction of weather observations would therefore be improved by the introduction of a

method in which the cyclonic and anticyclonic weather elements shall be distinguished in accordance with a natural and simple classification, and reduced in such a manner as shall indicate the prevalence of one kind of weather or another more distinctly than is now done in the customary averages and summaries.

For example, a curve of temperature, such as is made by a thermograph, exhibits a general rise and fall of cyclonic period, in addition to the more rapid rise and fall of diurnal period. If a pair of lines is drawn tangent to the curves of diurnal range, one above and the other below the temperature curve, the space between them may be called the temperature belt. A curve through the middle of this belt will represent the cyclonic range of temperature. The range is stronger and of shorter period in winter than in summer; its form is often unsymmetrical, showing a more rapid fall than rise. In winter, the diurnal range is small during the cyclonic fall; and larger during the cyclonic rise. In summer, the diurnal range is least during the prevalence of cyclonic clouds. The form of the diurnal curve is distinctly different during the prevalence of different winds. The facts here referred to, and others of similarly irregular period, have much to do with determining the character of the weather that we suffer from or enjoy; and they deserve as careful recognition as others which are customarily considered.

318. Weather Bureau of the United States. Soon after the whirling and progressive motions of cyclones were recognized and their control over weather changes was perceived, various projects were devised for the prediction of the unperiodic weather changes. The essentials of all these propositions were:— the appointment of a number of observers at selected stations; the uniform observation of the weather elements at the same moment of time once, twice, or three times a day at all the stations; the telegraphic transmission of the observations to a central station; the charting of the data thus concentrated; the forecasting of coming weather changes for different districts by inference from the charted weather; and the distribution of the forecasts by telegraph to the newspapers, or to numerous stations where weather signals might be displayed for the information of the public. Such a plan was gradually put in operation by Leverrier in France between 1855 and 1863; in the Netherlands in 1860; and in Great Britain in 1861.

In this country, Professor Cleveland Abbe, with the aid of the Cincinnati Chamber of Commerce, established a weather service on a small scale for the Ohio valley in 1869. This was followed in 1870 by a much more extensive scheme, developed by General Myer, Chief Signal Officer of the Army, and adopted by Congress as a national service. A corps of weather observers was then established all over the country. Professor Abbe was retained as scientific adviser at the central office in Washington, where he was for some

time the only predicting officer on duty. The service was greatly extended from year to year under General Myer, and after his death by his successors, Generals Hazen and Greely. On July 1, 1891, the meteorological division of the Signal Service was by order of Congress transferred from the Army to the Agricultural Department, under the title of the Weather Bureau, and Professor M. W. Harrington was appointed as its chief.

There were 24 observing stations in 1870; in 1893 there are 136, besides several stations of the Canadian Weather Service from which daily reports are received. Observations were first taken and synoptic charts prepared three times a day; but on January 1, 1889, two observations a day, at eight o'clock, morning and evening, were substituted. As at present developed, after continued growth for over twenty years, the Weather Bureau has a central office in Washington, with a staff of predicting officers and a number of assistants for the discussion of its voluminous records. There are stations in all the larger cities, at which observations are made systematically on pressure, temperature, wind, humidity, sky, precipitation, etc. These observations are corrected for instrumental errors, and the barometric readings are reduced to sea level (Sect. 107). The records are then translated into an abbreviated cipher and telegraphed promptly to Washington; the messages having precedence over all others. They are generally all received in the Washington office within an hour of the time of observation. The cipher is then translated back into the original form, and the data are charted on a number of maps. The predicting officer on duty for the time then draws in the isobars, isotherms, lines of equal change of temperature, etc., and at once proceeds to prepare a written statement of the probable condition of the coming weather for different districts of the country.

The difficulty of this work is very great. It must be done rapidly; an hour and a half or two hours from the time of observation must ordinarily suffice for its completion. The weather for forty-five divisions of the country must be determined separately; and in every case the predictions must cover a definite number of hours; ordinarily twenty-four but sometimes thirty-six hours from the time of observation. Besides the general forecasts for publication in the newspapers, special predictions are often called for, such as warnings of frosts, river floods, cold waves, off-shore winds, or expected dangerous storms. Several of these involve special instructions to stations within a certain district for the display or withdrawal of signal flags, in accordance with which vessels will delay their departure from various ports, or set sail on their voyages. The burden of these manifold duties on the predicting officer is so great that in recent years the experiment has been tried of appointing a number of local forecast officials in the chief cities, to the number of twenty-six, and supplying them with telegraphic data sufficient to construct synoptic weather maps and prepare local forecasts for a neighboring

region. There does not, however, at present appear to be a great increase in the accuracy of local predictions over those for the same period and district issued from the Washington office.

Besides the regular observing stations of the Weather Bureau, there are numerous stations on the sea coast and the Great Lakes, where warnings are given of approaching storms. Weather flags and other signals are displayed by volunteer assistants of the Bureau at numerous points in all parts of the country, for the benefit of farmers and others. In addition to the regular paid officials of the Weather Bureau, there are some two thousand voluntary observers in different parts of the country, who report through the state weather services or direct by mail to Washington once a month, and whose records are employed with those from the regular stations in the determination of climatic data. Voluntary observers are furnished with a pamphlet of instructions, and are in some cases supplied with instruments. At different times, special investigations have been undertaken with the assistance of voluntary observers; thunder storms and tornadoes being the most generally interesting subjects studied in this way.

The two daily weather maps, issued from the central office in Washington and from over seventy other cities in different parts of the country, are supplied to persons who will expose them conveniently for the information of the public; they are also sent in large numbers to schools. The daily edition of these maps, outside of Washington, is about 8,000. A Monthly Weather Review is published and distributed to all coöperating observers, giving a great body of information from regular and voluntary reports; with an account of the storms, winds, temperature, precipitation, etc., of the month, and with charts of cyclonic tracks, rainfall, temperature, and so on. Weather crop bulletins are issued every week during the growing season, giving information of interest to farmers.

In addition to the daily maps of the Weather Bureau, mention should be made of the Monthly Pilot Charts of the North Atlantic Ocean, issued by the Hydrographic Office of the Navy Department at Washington primarily for distribution to masters of vessels; and containing a large amount of generalized and current information concerning the winds, storms, ice, etc., of the North Atlantic from month to month. All notable storms are carefully followed between North America and Europe by means of reports from vessels; and in several cases especial accounts of them are issued as supplements to the Pilot Charts.

319. Weather maps. Of all the publications of the Weather Bureau, the daily weather maps are of the greatest interest to the student of meteorology, whether a beginner or an advanced investigator. If the facts commonly shown on these maps are not familiar from school years earlier than those in which

this book is employed, the following plan of studying them may be introduced; it being advisable that the study of the maps should proceed as a course of "laboratory work" parallel to the study of the text.

The meaning of the conventional signs on the map should be first learned.[1] Practice in drawing isotherms and isobars of simple examples should be afforded. Verbal descriptions of various weather elements should then be attempted, as: "the temperature is low in the northwest and high in the southeast," or "a large cloud area covers the Ohio valley," and so on; each element being considered separately. By shading the areas of different temperatures or

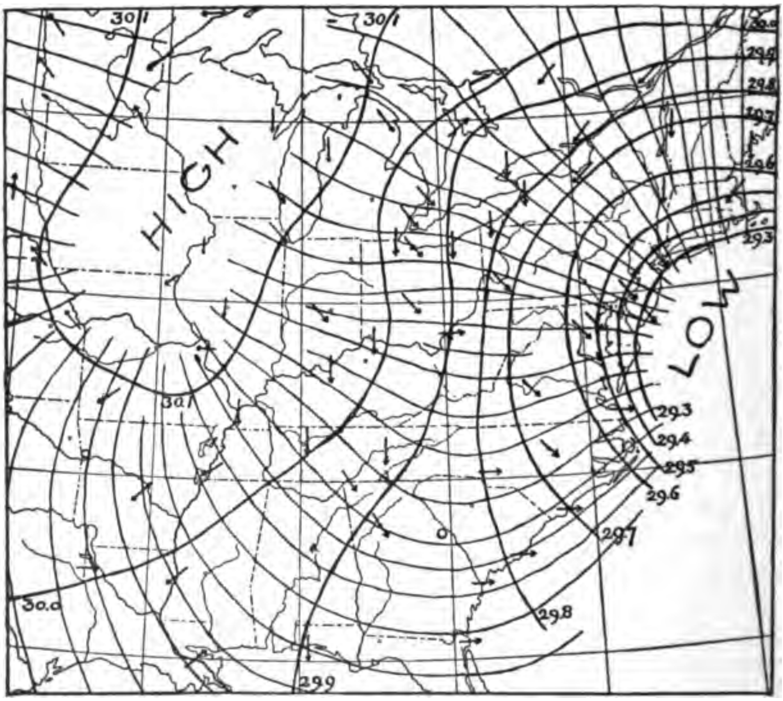

FIG. 106.

pressures, an instructive series of maps may be prepared for school use. Besides the lines of equal temperature and pressure, called isotherms and isobars respectively, lines of the most rapid decrease of temperature and pressure (at right angles to the isotherms and isobars) should be drawn on trial maps, and their course described. The lines of decrease of pressure will be frequently seen to diverge from areas of high pressure, or to converge toward

[1] In elementary teaching, the reduction of barometric observations to sea level cannot be explained; it must suffice to say in effect: certain corrections are applied to the observations at different stations in order to make them comparable.

areas of low pressure. The rates at which the temperature and pressure decrease along their respective lines — the gradients of temperature and pressure — should be determined, and expressed numerically and verbally. The isobars and pressure gradients are drawn in Fig. 106 for 8 P.M., November 28, 1888, the date of a disastrous cyclone on the Atlantic coast.

As the wind arrows represent the movement of the air only at isolated points of observation in great atmospheric currents that sweep broadly over the country, additional wind lines should be added between the arrows for the more graphic illustration of the presumed facts, as in Figs. 53 and 54. The greater or less velocity of the winds, indicated by figures near the arrows on the published maps, may be graphically illustrated by heavier or lighter wind lines. The inflow and outflow of great eddies of wind — cyclones and anticyclones — should be discovered, and the prevailing wind velocities for each recognized. Outline maps for characteristic weather types should be prepared, like those given in Figs. 74 to 79.

After deliberate practice in describing the several weather elements separately, their correlation should be considered. The most important correlation is that between the rate and direction of pressure decrease, or barometric gradient, and the velocity and direction of the wind. The general deflection of the wind to the right of the gradients and the increase of the velocity on stronger gradients will soon be perceived. In more careful study, a numerical relation may be established between these paired factors. By tracing the wind arrows from various maps on a single sheet of transparent paper, whose center is always placed over the center of low or high pressure, with one side always turned to the north, a "composite portrait" of a large number of winds in their proper relation to the low or high pressure center may be obtained, from which their inward or outward spiral courses will be at once perceived. The distribution of humidity, of clouds, and of rain or snow, and the peculiar warping of the isotherms with respect to areas of high and low pressure, are of equal importance. When it is perceived that the winds flow obliquely towards a center of low pressure, their escape upwards must be inferred; and the prevalence of clouds and rainfall in such regions may then be associated with the adiabatic cooling of the obliquely ascending currents. The opposite relation will be inferred in areas of high pressure.

After perceiving the systematic correlation of the various weather elements with the centers of cyclonic and anticyclonic areas, the movement of the center of these areas may be traced on the maps of successive dates; their paths may be gathered on a single map at the end of a month, and their average velocity and direction determined. In as much as the association of weather elements with these high and low pressure centers is relatively independent of their position on the map, a typical diagram of the weather around a cyclone or anticyclone, with its winds, temperature lines, clouds, and so on, may be

prepared on a sheet of tracing paper on the same scale as the maps, and by moving this diagram across a map along an appropriate track, the succession of weather changes experienced at any point on the map beneath it may be simply illustrated. All the phenomena of veering and backing winds, of rising and falling temperature, and so on, may be thus made as plain as desired.

During this advance in the study of the weather maps, there should be frequent local observations of the weather, and the facts thus noted should be compared with the larger collection of facts from many stations on the corresponding weather maps. The correlation of local and general weather will thus be clearly perceived, the essential facts being discovered gradually by the members of the class rather than taught by the teacher. Practice in prediction may then be introduced; but this should not come too soon, for fear that it might deteriorate into guess-work. It is not a little interesting to remark how easily, rapidly and naturally a class of school children may, by means of the wonderful collection of facts simply presented on the weather maps, advance through a study which but a few score of years ago was open only by laborious investigation to the most distinguished meteorologists of the world.

320. Methods of weather prediction. The essential principle employed in the prediction of weather changes by means of weather maps depends on the general eastward movement of weather areas. If a center of low pressure is charted in Colorado, while an area of high pressure stands over West Virginia, southerly winds and rising temperature will be predicted for the area occupied by northerly winds or calms east of or within the area of high pressure; rains will be announced for the area about Missouri over which clouds are forming east of the cyclonic center; and clearing weather, westerly winds and falling temperature will be inferred for the cloudy and rainy region of the cyclone itself. Supplementary to this main rule, there is a general tendency to an increase in the intensity of storms as they approach the Atlantic coast. Variations from the average direction or velocity of progress of weather areas are seldom accurately foretold. When the distribution of pressure is equable and no distinct areas of high or low pressure are contained within the limits of the map, every predicting officer may have his own method, more or less consciously defined, of determining the most probable sequence of weather. At times of decided departure of any element from the normal, a return toward the normal is usually predicted, even if there is no very apparent reason for it: for this purpose, charts have been published by the Weather Bureau giving the normal values of certain weather elements for different stations and for short intervals of time, such as periods of ten days.

In reviewing this statement, and even while appreciating the great value of the predictions, one cannot avoid a certain feeling of disappointment that

all the labor that has been bestowed on the subject of weather prediction in this country during the last twenty years has not led to greater advances beyond the methods employed and the results gained at the outset of the undertaking. The number of stations has grown, and their equipment has been materially improved; the accuracy of various processes preparatory to charting has been increased; a vast body of information has been accumulated for study relative to the kinds and changes of weather; various predicting officers have had extended practice in their art; and while the forecasts are truly made for longer periods than they were at first, and are certainly superior in definiteness and accuracy to those issued twenty years ago, their improvement is not so great as was hoped for. Mistakes in prediction are still made, and of much the same kind as at the beginning of the service; it is still often impossible to predict the weather changes for more than twenty-four hours in advance with a desirable degree of certainty. It must be concluded from this that the limit of accuracy of weather prediction by present methods is practically reached; and that no considerable advance over the inherent difficulties of the subject can be expected until some distinctly new method of observing, charting or forecasting is introduced.

321. Distribution of predictions. On the other hand, a decided advance has been made in the distribution of the predictions and in their utilization by the people. Besides a very general publication of the forecasts in the daily papers, with an increasing completeness of statement year by year, there has been a remarkable growth in the number of daily weather maps distributed from many centers, and a gradual and general increase in the number of display stations of different kinds. A generation of our population has grown up, accustomed to look for and to use the forecasts of the Weather Bureau. Apart from the value of the forecasts to the general public, to whom they are a matter of convenience rather than a necessity, innumerable examples of their more technical value might be cited. Coasting and fishing vessels are informed before leaving port if they are likely soon to meet a storm on their course; or they are warned not to sail if dangerous winds are immediately expected; and this information is particularly important to the smaller craft engaged in the local coasting trade. Wrecks are relieved by assistance summoned by messages from the coast telegraph stations. Disasters on the Great Lakes have been materially diminished by warnings of coming storms. Proper care is given to various crops at critical times in their growth or harvesting; although in this direction, the farmers of our country have yet much to learn.

322. Weather and storm signals. The flags now employed for indicating the predicted weather are as follows: — a square white flag for fair weather; a square blue flag for rain or snow, no distinction being made as to amount;

a triangular black flag for temperature, indicating warmer weather, if displayed above the white or blue flag; colder weather, if below. The changes of temperature thus indicated are always to be understood as being independent of diurnal changes; and hence show an expected rise or fall compared to corresponding hours on the preceding day. A square white flag with a square black center indicates the coming of a cold wave. This is not displayed unless it is expected that the temperature will fall suddenly and decidedly to below 42°, and generally to a much lower temperature; twenty-four hours or more notice is generally given of this change. Similar signals are displayed on either side of the baggage car on railroad trains in some parts of the country; steam whistles are also employed, a system of long and short blasts sufficing to express the predicted changes.

Storm signals displayed at ocean and lake ports are as follows:—The cautionary signal, displayed only on the Great Lakes, is a square red flag with white center; this indicates the approach of winds not so severe but that sea-worthy vessels can meet them without danger. The storm signal, indicating severe winds, is a square red flag with a black center. A white or red pennant, displayed with the cautionary or storm signal, indicates the direction of the expected winds: a red pennant above the square flag is for northeasterly winds; below, for southeasterly winds; a white pennant above the square flag is for northwesterly winds; below, for southwesterly. The red pennant alone indicates that the local observer has received information from the central office at Washington concerning a storm that may prove dangerous to departing vessels. At night a red light indicates easterly winds; a white light above a red indicates westerly winds.

Many instances might be quoted in illustration of the manner in which weather changes have been heralded by predictions and signals in season to allow all possible preparation to be made for their coming. Thus, a cold wave is perceived while yet in its earlier stages as an accumulation of cold air under high pressure in Winnipeg, north of an advancing cyclonic center in Texas or Arkansas. The wave is then announced as a winter "norther" before it reaches the Gulf coast; its further progress eastward is anticipated by information sent to the cities in the Mississippi valley; its extension up the Ohio valley and over the southern states even as far as Florida and to the middle Atlantic seaboard is fully predicted. Stock raisers on the western plains, railroad employees in charge of cattle trains, beef companies and pork packers in the central states, shippers of perishable goods, ice companies waiting to gather their winter crops on the northern lakes or rivers, and orange growers in Florida, all prepare for the best or worst that the cold wave may bring them. The janitors in large buildings in our cities strengthen their fires when the cold-wave flag is displayed, and the insurance patrols redouble their vigilance in anticipation of conflagrations so often caused by

overheated chimneys, and so difficult to extinguish when a gale fans the flames and water freezes on the walls.

In a similar manner, the arrival of a tropical hurricane at Cuba is announced by submarine telegraph, and its deliberate but destructive advance along our southern Atlantic coast is duly published. The late frosts of spring and the early frosts of fall are foretold to fruit farmers; cotton planters in the south look for warnings of wet or damp weather. The news of the ending of a hot wave in summer by a turn to westerly winds, or of the end of a drought by the advance of rain storms, is watched for by thousands or even by hundreds of thousands in city and country. The more intelligent the population, the greater is the use made of weather forecasts.

Unhappily, storms and frosts cannot always be correctly foretold; and the changes that are announced do not always come to pass. Exceptional atmospheric movements cannot be predicted by the methods now employed.

323. State weather services. Although the records of state weather services are useful chiefly in climatic studies, yet their present close association with the national Weather Bureau suggests their description at this place. Half a century ago, systematic observations were undertaken by unpaid observers in New York and Pennsylvania; but these were discontinued after a few years. The earliest state service in recent years was established by Professor Hinrichs in Iowa in 1873. At first neglected by the national service, the state services were later greatly extended by its aid, and since the transfer of the Bureau to the Agricultural Department, they have been carried into all the states and territories. In some cases they are supported by small sums of money annually granted by the local legislatures; they are then well equipped with uniform and accurate instruments, and their records are fully published; but the work of the observers is always gratuitous. In other cases no state support has been granted, and the observations are less systematically carried on. In all cases, a member of the national service is detailed to supervise the work of the local volunteer observers; some form of publication is maintained, and the digested results of every month are early forwarded to Washington for use in the preparation of the climatic tables and charts in the monthly Weather Review. The observations required are all of the simplest character, being limited as a rule to temperature aud precipitation; winds and clouds are sometimes included. Occasionally, studies of subjects a little aside from the usual routine of weather observations have been attempted, and it is extremely desirable that such investigations should be extended. Even where the most work of this kind has been done, there is a wide field open for new or better work.

It is very important that the observers in the state services who are willing to give their time perseveringly and regularly to the long task of determining

local climatic data, should be equipped with good instruments, and not waste their efforts in making records whose accuracy is not proportionate to the care bestowed on them. The moderate cost of good instruments in the beginning should be somehow met; otherwise it is hardly worth while to begin the task. Once well begun, it should be persevered in; for the value of local records increases greatly as their uninterrupted duration is prolonged. Whenever possible, an analytical account of local weather changes, a comparison of local events with the general phenomena of the weather maps, and a careful discussion of the results should be attempted, even if only for private information, somewhat as indicated in Section 317.

324. Private meteorological observatories. In a few instances, private observatories have been established for the study of meteorology. The most noted of these in this country is the Blue Hill Observatory, established near Boston by Mr. A. L. Rotch in 1885. A few years after its foundation, it was associated with the Astronomical Observatory of Harvard College, in whose Annals its records are elaborately published. Notable among these is a series of cloud measurements, the most extensive yet undertaken in this country. Figs. 65, 66, 68, 69 have been prepared from these records by the observer, Mr. H. H. Clayton. The establishment of similar observatories in other parts of the country would do much for the advance of the science.

325. Foreign weather services. Since the establishment of the first weather service by Leverrier in France, similar services have been organized by nearly all the civilized nations of the world. The following countries publish weather maps and issue forecasts:—Great Britain, France, Germany, Belgium, Austria-Hungary, Switzerland, Italy, Russia, Spain, British India, Japan, Australia, and New Zealand. Canada and Cape Colony issue forecasts, but do not publish maps. The following countries maintain a system of observations, and issue certain reports, but do not prepare daily maps or issue forecasts:—Norway, Sweden, Denmark, Netherlands, Portugal, Roumania, Algeria, Mexico, Brazil, Argentine Republic, Chili, China, and certain smaller countries.

With the exception of India, which lies largely in the torrid zone, the features of the weather maps of different countries are astonishingly uniform in general features, which differ for the most part in intensity and in frequency of occurrence. The wide-spread distribution of cyclonic and anticyclonic disturbances, their prevailingly eastward movement around one pole or the other, the control of weather changes by these drifting centers of low and high pressures, and the relation of smaller storms to the larger ones, have been completely established by this vast fund of original and definite information.

Daily weather maps for the North Atlantic Ocean, already referred to in Section 232, have been published for the year beginning August, 1882, by the British meteorological council; and for subsequent years jointly by the German and Danish weather services. An International Bulletin was published for a number of years by our national Weather Service, including weather maps for the entire northern hemisphere. The data for this extensive undertaking were secured from the weather services of various countries, from numerous naval and mercantile vessels, and from certain other observers. The chart of circumpolar storm tracks, Fig. 62, was prepared by Loomis from the maps in these Bulletins.

326. Weather proverbs and weather lore. There is a great number of popular sayings concerning the weather; some being of value, others deserving only to be classed with the superstitions of the middle ages concerning comets and shooting stars. A few of these may be considered. In our latitudes, a fine day is often called a weather breeder, and is said to be too good to last; thus recognizing the generally rapid succession of bad weather after good. In the same way, it is sometimes pronounced to be too cold to snow during cloudy weather in winter; the cold resulting from a preceding anticyclonic calm not sufficing to produce snow until a southerly wind springs up in front of the next cyclone and raises the temperature. The increase of humidity in the cooling southerly winds in front of a cyclone gives rise to many prognostics. Rheumatic pains increase; houses with stone walls, being in summer somewhat cooler than the outer air, become extremely damp, the walls even dripping wet, when the air becomes sultry before a rain; smoke falls before a storm, not because the air is then lighter from decrease of pressure, but because the condensation of vapor on the smoke particles weighs them down. Dew formed plentifully after a fair day and soon dissolved the next morning indicates strong range of temperature under the clear sky of an anticyclone; and hence may be taken to foretell a day or two of fair weather, followed by a cyclonic area.

The proverbs relating to the winds are very numerous. The immediate accompaniments of the winds are expressed when we say a southerly wind brings rain; a northwest wind brings cooler or colder weather; this being dependent on their position in cyclonic areas. Of course, these rules have to be reversed in the southern hemisphere. In Scotland it is said "the northwest wind is a gentleman and goes to bed"; meaning that the nights are calm after northwest wind by day; this follows naturally from the diurnal variation of velocity in the clear weather of such a wind. The changeableness of weather in the temperate zone is expressed by the sailor's saying: "a nor'wester is not long in debt to a sou'wester"; the two being separated only by the anticyclonic area or ridge of high pressure. The prevalent path of our cyclones is north of

the regions of greater population both in America and Europe; hence the usual sequence of the winds in clearing weather is from the southeast through the south to the west; and this is called veering with the sun. If on reaching the west the wind backs through the south towards the east, another cyclone is coming with its spell of bad weather. Hence the saying: "when the wind veers against the sun, trust it not, for back it will run." This saying is sometimes applied to the case of storms that clear by the backing of the wind from an easterly quarter through the north to the northwest, as happens when the storm passes south of the observer: "Back it will run" is then taken to mean that another storm is coming; but here the saying does not appear to have good justification, for it is plain that if two equally competent observers, one on the right and the other on the left side of a cyclonic path, should employ this rule to guide their forecasts, they would be in continual contradiction. In northern Canada, the backing of the winds through the north must be the ordinary occurrence. It may be, however, when a storm passes us to the south, having come along the coast from the Gulf of Mexico, and thus giving backing winds as its clouds clear away, that another cyclone from the northwest may soon follow, and thus give some countenance to the saying.

Prognostics from fog and clouds are of much value. The formation of fog in valleys at night, and its dissipation early the next morning indicates fair weather for a time; for this implies clear anticyclonic air, with active radiation at night and warm sunshine by day. In the same way, the dissolution of cumulus clouds about sunset, only their upper parts remaining for a time as thin alto-cumulus layers, implies a diurnal control of the weather, and hence a fine morrow, under anticyclonic pressures. Cirrus clouds generally indicate the approach of bad weather. "Mare's tails and mackerel scales make lofty ships carry low sails"; these thin clouds being the elevated overflow of an approaching cyclone. At the same time, the sun and moon are paled by the cirrus veil; they are frequently surrounded by halos formed by refraction in the ice crystals of such clouds; the course of the high cloud streaks is often strongly different from that of the surface wind; and all these signs forebode a change towards foul weather. As the clouds gather in lower levels, the light reflected at night from iron furnaces and from the electric lights of modern cities, is seen with increased brightness, and indicates the speedy approach of rain. As the storm clouds pass by, a break in the cloudy sheet showing enough clear blue sky "to make a Scotchman's jacket," or "a Dutchman's breeches," as it is variously expressed, shows the coming of fair weather; for while breaks may often occur in one cloud layer or another within the stormy area, it is very seldom that clear blue sky can be seen through such spaces; but in the rear of the storm, when the lower clouds are gone, and the high cirro-stratus sheet remains projecting backward from the storm center, but drifting along with it, a break discloses the bright blue sky above.

The colors of the sky may be often used as prognostics. A clear, fresh blue shows the approach or presence of an anticyclonic area, while a pale sky forebodes an approaching cyclone, even before its cirrus streamers appear. Halos are commonly formed around the sun or moon in the thin cirro-stratus clouds before a cyclone. A glaring, hazy sky often comes with southerly winds and increasingly hot weather in summer; little dew is then formed at night, as radiation is checked; frosts need not be expected at such times. A clear stretch of sunset red close along the horizon, surmounted by yellows, indicates fair weather the next day; and especially so if the rosy segment is well displayed above the horizon colors; but a lurid western sky at sunset, with colors spread above the horizon on thin cirrus clouds, indicates a coming storm; and if the sunset be dull and "dirty," with clearer sky in the east, the storm is nearer. Rainbows in the east and hence in the afternoon, foretell clearing weather; these bows being generally formed on the rain of retreating thunder showers (Sect. 290); but if seen in the west and therefore in the morning, rain is approaching. This of course applies only in those latitudes where thunder storms move from west to east.

A variety of sayings relate to the behavior of wild and domestic animals. Some of these depend simply on their behavior as affected by the condition of the weather at the moment, and such may be frequently relied on; beasts and birds as well as man being disturbed particularly by changes from dry to damp air; but there is no proved value in the sayings which attempt to foretell the character of the coming season by the supposed instinctive foresight of such animals as bears, beavers, moles and squirrels, in the preparation for severe winters or long droughts. Actual investigation has shown that these dumb animals have no such foresight as is popularly attributed to them.

No credence should be attached to the innumerable sayings regarding the character of certain seasons as determined by the weather on certain dates of the calendar; a careful and extended account over a number of years would undoubtedly show as many failures as successes in such predictions; that is, the reputation of such weather proverbs comes only from "counting the hits and forgetting the misses," as is so common in careless generalizations. The same comment may be made regarding the days of the week in which the phases of the moon change, and the attitude of the new moon in the sky. It is manifest that the tilting of the horns of the new moon, for example, will be the same for all observers on a given latitude circle; and that for a given number of days after new moon, these observers will have a great variety of weather; yet this lunar prognostic would give them all the same.

Nothing but a continued statistical study of prognostics, and a strict comparison of forecasts with facts, will suffice to demonstrate their value; yet no such comparison has been made by the greater number of persons who credulously give faith to oft-repeated sayings, believing that there must be some-

thing in them, because they are often said. They should be regarded as survivals of superstitious folk-lore, rather than as weather-wise sayings.

327. Weather cycles. Many efforts have been made in this country and abroad to discover a periodic recurrence of weather changes of longer and more regular intervals than those between successive cyclonic centers. A weekly recurrence of similar conditions has long been known in a general way, but the conditions which determine its duration and its variations from perfect periodicity have not been discovered and it has seldom been made practically useful. It appears to depend on a double cyclonic period. The control of the weather by the moon has long been a favorite idea, but it has not been found to bear the test of accurate comparisons of weather and lunar phases, except in a very faint and imperfect manner. Whatever slight excess of one weather element or another there may be at certain times of the lunation, they have no sufficient value for use in weather prediction. Thunder storms in Europe have been found slightly more frequent at the time of new moon and first quarter; but not to a sufficient degree to warrant the use of these poorly marked cycles in forecasting. A period corresponding to that of the sun's rotation, or 26.7 days, has been found to accord with a faint variation in the intensity of various weather elements; and it is argued that this indicates an effect on the atmosphere produced by solar magnetism as well as by solar heat. Besides these periods of an astronomical nature, there are others of apparently arbitrary duration, probably corresponding to the period of the surges already referred to (Sect. 105). These have not been sufficiently studied to know how far they may be useful, or to determine their cause; but they appear to be deserving of study.

The control of the weather by the moon or the planets still occasionally finds enough believers to support the publication of elaborate long-range weather predictions. As these are couched in general language and intended to be applicable to large areas of the country, it is not at all difficult to gather a number of verifications for them; but they are no better than the forgotten predictions of astrology of centuries ago.

CHAPTER XIV.

CLIMATE.

328. Climate. The average values of the atmospheric conditions of a region constitute its climate. The most important climatic elements are first, temperature; second, various forms of moisture, as vapor, cloudiness, and precipitation; third, wind, including storms. The pressure of the atmosphere is not a climatic element, and needs to be considered in this chapter only in its association with the divisions of the wind system.

While annual averages were first considered in the definition of climate, more and more importance has come to be attached to the average of seasonal values; and to such special quantities as the average highest or lowest temperature or rainfall of a season or a month. Even the extreme values are often included in climatic tables, in order to present as fully as possible the meteorological features of a district; but in so doing, we approach the consideration of its weather. A full climatic account of a locality should include: for temperature — the monthly and annual means, the mean diurnal range for the several months, the mean and the absolute extremes for the year and the months, the mean diurnal variability (the mean of the differences between the successive daily means), the average dates of latest and earliest frost, the average number of days without frost; the average duration and value of cyclonic ranges of temperature in the several months (Sect. 317); the mean intensity of sunshine in clear weather of the different months; the mean temperature of the soil at successive depths down to five or six feet: for moisture — the monthly mean absolute and relative humidity, the mean monthly evaporation from a water surface; the mean cloudiness and mean duration of sunshine in the several months; the mean monthly and annual rainfall, with additional data for melted snow in the winter months; the mean number of rainy and snowy days in every month, the mean frequency of rainfall in every month (number of rainy days divided by the total number of days), the average dates of latest and earliest snowfall, the average depth of snow on the ground at the end of every month; if possible, the proportion of rainfall received from general cyclonic storms and from local thunder storms in the several months, and the mean diurnal variation of rainfall for the different months; the mean number of days with thunder storms and with hail in the several months: for winds — the frequency of different directions for the several months, with the corresponding mean velocities, and indication of the frequency of calms and of exceptionally strong winds; the mean diurnal variation in direction and velocity for several months.

In a region like the eastern United States, the means of climatic elements in corresponding months of successive years vary so greatly that a considerable number of years is required to determine their true values. Hence the importance of maintaining weather records continuously under conditions as nearly constant as possible, in order to outlast the influence of dry or wet, warm or cold periods. As has already been said, it is hardly worth while to begin such records unless there is a fair probability of their continuance, and unless good instruments can be secured and properly exposed.

329. Climatic zones and subdivisions. As it is impossible to describe the climate of every locality in the world, even if it were known, it is customary to class together certain large areas over which the climatic conditions are similar. The earliest attempt of this kind, originating with the Greeks, considered only the divisions of the earth as theoretically determined by the geometrical distribution of sunshine; yet this suffices so well for the purposes of elementary description that it is still followed in the usual accounts of the torrid, temperate, and frigid zones, whose definition by latitude circles has already been given in Chapter III.

An inspection of the isothermal Charts I, II, and III, suffices to show that the regular boundaries of the zones are divergent from the undulating lines of equal temperature; and hence some authors propose to limit the zones by certain selected isotherms. This plan has much to recommend it, but it is unsatisfactory in giving undue prominence to temperature alone, and neglecting sufficient consideration of other climatic factors. It is therefore here proposed to consider climatic belts as limited by the wind systems rather than by the isotherms or the parallels, and thus secure a closer accordance with natural atmospheric areas. The limits thus determined do not differ seriously from the lines of latitude; they are necessarily somewhat indefinite, but in this they accord with the gradations of nature, where sharp dividing lines are not drawn.

There would thus be recognized a torrid zone, extending somewhat beyond the tropical circles to the margins of the trade wind belts; through the middle of this zone runs the equatorial belt; on either margin lie disconnected subtropical areas. The temperate zones then follow over the latitudes controlled by the prevailingly westerly winds. They are somewhat narrowed from the breadth they would have when defined by the tropical circles; they merge into the frigid zones, from which the polar circles may serve to separate them. The temperate zone is not well named, as its actual variations of temperature are very strong over large areas.

From what has preceded in earlier chapters, it is manifest that the zones are by no means of uniform character over their entire area. The north temperate zone in particular includes areas so diverse climatically that no simple description can be given of them all. A subdivision of the zones

according to land and water areas is therefore required; and these must again be subdivided into interiors, western coasts and eastern coasts. Finally, in each zone, the striking differences determined by altitude above sea level will deserve consideration under a heading of mountain and plateau climate. If this subdivision be regarded as too complex, a return may be made to the simpler division by latitude circles; but it should be remembered that effort should be made to recognize and classify natural features, rather than to force natural features into an arbitrary classification. In all cases, the general characteristics of each zone and of its subdivisions should be considered, rather than their limits, which are necessarily belts instead of lines.

330. The torrid zone. The margins of the torrid zone, as ordinarily defined, lie in latitude 23½° on either side of the equator. When defined by isotherms, they are generally placed on the lines of 68° or 70°, whose course may be traced on the chart for the year, widening on the continents and narrowing eastward across the oceans; corresponding nearly with the polar limit of palms. When defined by the polar margin of the trades, the zone lies between latitudes 30° and 35° north and south, widening somewhat on the eastern side of the oceans. The area of the zone according to the latter limits greatly exceeds that of the former.

The chief feature of the torrid zone according to all these definitions is a prevailingly high temperature. The special feature of the first definition is the small variation of insolation during the year; according to the second, the prevailingly high mean annual temperature; according to the third, a comparative constancy of weather through the year, in which the simplicity of the climatic features is apparent. Each tropical circle crosses the broad trade wind belts, and thus separates these areas of extraordinarily uniform features into two parts: it therefore seems advisable to abandon these limits and to widen the torrid zone until it shall include the whole belt of steady winds and simple climate. The isothermal limits of the torrid zone narrow it on the eastern side of the oceans, where the action of the winds tends to maintain a constancy of seasons, and broaden it on the western side of the oceans, where a greater variation of weather and a correspondingly increased complexity of climate is caused by the extension of continental conditions over a marine area; it therefore again seems better to adopt another division whereby the equable and mild climates of the eastern ocean margins even to latitude 35° may be associated with the great uniform torrid zone, and by which the more variable eastern coasts of China and the United States may be associated with the variable north temperate zone.

331. The oceans of the torrid zone need little additional description after what has been said of their several climatic features in the chapters on tem-

perature, winds and rainfall; and on the succession of phenomena in the chapter on weather. The temperatures never reach extremes, unless in the lee of a large land area, as to the west of Africa, where the winds sometimes carry hot desert air in summer or cool desert air in winter over the sea. Were it not for the excessive humidity by which the discomfort of the equatorial belt is produced, the ocean area of the torrid zone would deserve the name of temperate better than the northern zone to which it is commonly applied. Its winds are of unequalled steadiness; their interruption by cyclones being altogether exceptional. Along its equatorial portion, the mean annual temperature is moderately increased, cloudiness and rainfall are decidedly increased, and the winds are weakened.

332. The lands of the torrid zone. The torrid lowlands, outside of the sub-equatorial belts, are in great part desert areas, or but scantily clothed with vegetation; not because of unfit temperatures or barren soil, but because of deficient rainfall. The Sahara in one hemisphere and central Australia in the other give the most extensive examples of the trade wind desert. On the western coasts, the deserts come directly to the sea shore. It is within these arid areas that the extremely high temperatures of the globe are reported. With a mean annual temperature about 80, the mean of the mid-summer month rises above 90°, with extreme temperatures of 110° or 120°. In the mid-winter month, the temperature averages about 70°, with minima seldom as low as 50°. The winds are especially steady during the latter season; their constancy being interrupted chiefly by their regular nocturnal cessation, at once a striking feature of both weather and climate in torrid deserts. The excessive dampness of the equatorial doldrums is here replaced by an excessive dustiness of the air.

The rainy equatorial belt on land determines the extension of the great equatorial forests of middle Africa and of the Amazon. Here the heat of the hot season is not so excessive as on the dry deserts some distance from the equator; here is the luxuriance of life which we commonly associate with too large a part of the torrid zone. Fortunately for the world, the breadth of the rainy belt on the lands is increased over its narrower limits on the ocean, and thus the area of the adjoining trade wind deserts is encroached upon. As the rainfall of the sub-equatorial belts decreases, so the vegetation diminishes; dense jungles of the equator give way to more open forests; and these in turn to a region of scattered trees with intervening grassy spaces; further on, the trees disappear except along the rivers, the grasses become sparse and patchy; then only a desert flora can survive the long dry season, and finally even this is lost, leaving the ground barren.

The Indian monsoon region has a climate of its own. Here the division of the year into three seasons is the most marked feature; a cold season, when

the normal trade wind holds sway, with little precipitation except in the more northern part of the peninsula; a hot season, when the northward march of the sun raises the temperature to an excessive degree; a wet season, when the winds weaken and change, and the moist southerly monsoon blows from the sea over the land with heavy clouds and rain. This is like the Saharan climate, with the addition of the equatorial climate for a part of the year. Northwestern India, known as the Punjab, or Five-river district, manifests this succession of seasons with much distinctness, as appears from the following abstract from an account by a resident: From April to June, there is little or no rain; the west wind blows from the deserts of the lower Indus with a desicating, scorching heat, as if from a furnace. The thermometer in the shade may rise above 120°. The nights are relatively cool, and then the houses are opened for ventilation. It is only in the early morning that an enjoyable air is found for exercise. During the day-time, the houses should be kept closed. Vegetation withers; the grass seems burnt to the roots; the ground is hard and cracked. Before the wet season opens, the winds weaken and the heat seems even more severe than before. In the later summer, the rains come as cyclonic storms of increasing intensity, yielding a plentiful rainfall over much of the country. Trees burst into leaf a second time; the grass springs up and a vegetation is soon developed that can be scarcely kept within bounds. The breaks in the rains are oppressively hot and sultry; and for a time after the rainy season closes, the heat and dampness are excessive, making September the most unhealthy month of the year. In October, the northerly winds set in steadily, clearing the skies; the temperature is then pleasant, and by the end of the year the nights become cold. Rain falls in moderate amount from cyclonic storms which move eastward, bringing cold northerly winds after them, as if they belonged to the procession of storms in the temperate zone; thus suggesting that northern India, which is so truly tropical in its hot and wet season, partakes of the features of the temperate zone in winter. In February, there is a short spring, tempting a growth of vegetation; but this is followed by a rapid increase of temperature, and the hot season is again at hand.

Two subordinate divisions of the torrid zone remain to be mentioned: the litoral and the mountain climates. The torrid coasts are generally well-watered, and their diurnal temperature is modified by the sea breeze. Western coasts in trade wind latitudes form exceptions to this rule, as they are prevailingly barren; but from this exception there is again a departure in the Indian monsoon region, where the western coasts are better watered than the eastern coasts.

The mountain climate of the torrid zone is tempered by altitude. In equatorial Africa, mountain peaks ascend even high enough to hold snow the year round; in rising from the torrid forests around the mountain base, one

may pass through successive zones of vegetation until the cold of the upper air prevents all plant life. In Ceylon, the highest mountain bears certain plants like those of England, while the lowlands exhibit the greatest tropical luxuriance. On the plateau of southern India, 6,000 or 7,000 feet above sea level, the days are warm and the nights cool, but the mean temperature is moderate all the year round; even in the wet season the rainfall is not excessive, as the higher hills to the west receive the greater part of it. "Many an old Anglo-Indian, whom choice or necessity has led to fix his home in India, has found in these hills scenery as beautiful and a climate as enjoyable as any in the most favoured lands of the Mediterranean shores." The rainfall of the highlands also departs from the rule of the lowlands; for mountains often provoke precipitation that might not otherwise occur. Thus the highlands of Brazil cause the growth of many a cloud from which the rain-trail drifts across the adjacent valleys; in the absence of the mountains, the dryness of the interior plains would extend closer to the coast. Even the barren desert of the Sahara contains mountains that cause sufficient rainfall to support forests, and yield streams that wither as they descend to the lower country. The bountiful rainfall of the Malayan islands is due chiefly to their combination of equatorial, litoral and mountainous conditions. Their accessibility and their productiveness destine them to play an important part in the higher development of the world.

333. The transitional sub-tropical areas between the torrid and temperate zones are of much greater importance in certain longitudes than in others. On the lands around the Mediterranean, on the middle western coasts of North and South America, along the southern coast of Australia, and in South Africa, their characteristic features are distinctly brought out in the dryness and warmth of the nearly continuous fair weather of the summers, and in the cloudiness and fair supply of rainfall that come with the more stormy weather of the winters. These areas, at one season dominated by the inflow at the source of the trade winds, at another by the stormy winds of the temperate zone, alternately associate themselves with the zones that they adjoin. They are far enough from the equator to avoid the excessive heats of the torrid zone; and their situation with respect to land and sea prevents their invasion by the low temperatures of higher latitudes.

The climate of these areas is among the most delightful of the world. The attractions of the health resorts of the Mediterranean have long been well known. The equally delightful climate of southern California has in recent years gained a deserved appreciation in this country. Although in the same latitude as our middle Atlantic coast, it has none of our hot or cold waves; its skies are prevailingly clear, and even in its winter wet season, its rains are light. In regularity of succession of diurnal features, and in constancy of

mean temperatures through the year, it rivals the greater part of the torrid lands.

It does not appear advisable to continue the sub-tropical areas around the world. On the oceans, they may follow the belt of high pressures, in which the winds are weak, the skies prevailingly clear, and the air fresh. But the continental interiors on these latitudes possess seasonal variations so strong that they should be associated with the temperate zone; and the eastern coasts, even nearer the equator than the sub-tropical areas of the western coasts, are more naturally associated with the land areas on their polar and western sides. Their annual range of temperature exceeds that of the tempered western coasts; their rainfall occurs in summer rather than in winter, like that of the interiors. Thus Florida, lying five degrees south of southern California, will be grouped with the temperate zone, which widens eastward in crossing the continents, rather than with the sub-tropical areas, which widen in crossing the seas. Southern China, near the middle latitude of the Sahara, will be grouped in the same way.

334. The south temperate zone. The temperate zones of the two hemispheres are so unlike that they must be described separately. The great water zone of the southern hemisphere will sufficiently represent the smaller northern ocean areas; the broad northern continents will exaggerate the climatic features of the restricted southern islands. The southern temperate zone, lying beyond the axis of the tropical high pressure belt, is chiefly an ocean zone. It is characterized by prevailingly stormy westerly winds, prevailingly low temperature, and frequent cloudiness and precipitation, rather than by the seasonal variation of these elements. Owing to the absence of land barriers, the southern ocean currents wheel round and round their polar center, with few pronounced north or south deflections, such as occur in the corresponding latitudes of our hemisphere; hence the southern temperate zone is of extraordinarily uniform features all around its circuit. In spite of the strong variation of insolation with the southing and northing of the sun, the range of mean monthly temperatures from January to July is hardly greater than at many parts of the torrid oceans; so effective is the conservative action of the ocean waters, and of the fogs and clouds by which they are so generally shielded; but winter is the season of stronger winds and heavier precipitation. In those more detailed climatic elements, which take account of the monthly and inter-diurnal ranges of temperature, there is a great difference between the torrid and the south temperate zones, because the former is controlled so largely by diurnal processes, and the latter is so completely in the hands of cyclonic processes, as has been stated in the chapter on weather. The few land areas that interrupt the south temperate ocean zone are most inhospitable; not that their winters are exceptionally

cold, for they know nothing of the extremely low temperatures of northern continental interiors of corresponding, or even of lower latitudes; but that their chilling summers are so little warmer than their winters. The temperature finds its summer maxima about 40° and 50°. Snows are not uncommon even in January or February, when the weather should be mildest. There is therefore no mild season in which provision may be made for the remainder of the year, as there might be if the lands were larger. When the uninhabited island of South Georgia, with glaciers descending into the sea, is compared with middle England, to which it corresponds in latitude, the contrast is remarkable; but this is more because of the exceptionally favorable condition of England in the north temperate zone than of the peculiar quality of the South Georgian climate in the south temperate zone.

335. The north temperate zone is the great land zone of the world. It is of particular importance as the chief seat of modern civilization. Its oceans need little consideration, as they repeat the features of the south temperate zone, modified by an approach to the land climate on the western side of the oceans, and by the diverse courses of the ocean currents, especially on the eastern side of the oceans. On the broad northern lands, the variation of the seasons and the inconstancy of the weather become the dominating climatic features; and annual averages, which almost suffice to define the climate of the torrid oceans, are very misleading. The actual temperatures of the year reside longer near the temperatures of the hottest or coldest month than near the annual mean; hence the hot and cold seasons are strongly separated by relatively short warming and cooling seasons. The interior of the lands will be first considered. Western Europe and the eastern United States, standing in different relations to their continental centers, will be then taken as types of leeward and windward coastal areas.

The southern portion of the continental interiors attain truly torrid heats in their summer season; Arizona and Persia rival the Sahara in having mean temperatures for July above 90°. Their lowlands are deserts; their scanty rains come chiefly from showers that begin on the neighboring mountains, or from violent thunder storms that deliver a whole season's precipitation in a few hours. Their winters are cool, with a January mean of about 50°, and minimum often down to freezing. The northern part of the interiors have warm summers, with a July mean of about 60°; but toward the Arctic border they are extremely cold in the winter season, when the January mean over large districts is as low as −20°; while in a limited district in northeastern Siberia it sinks below −40° or −50°; and in the associated margin of the frigid zone, at the town of Verkoyansk, even to −60°. In these hard frozen regions, the underground temperature remains below the freezing point the year round, trees are stunted or absent, and crops grow only in the surface

layer of soil that melts under the sunshine of long summer days. Over the greater part of these extended land areas, the rainfall has a more or less distinct maximum in summer; in the far interiors, the winter snowfall is moderate in spite of the severe cold.

A large interior area of North America, stretching from Missouri northward past Winnipeg and eastward beyond the Great Lakes, possesses in a marked degree the variable features of this so-called temperate climate. The summers are warm, with spells of extreme heat often broken by destructive local storms from which the greater part of the rainfall is obtained; the winters are cold, with times of excessively low temperature, brought by stormy winds that cause rapid changes from one extreme to another; vast floods of freezing air from the north are the scourge of the winter, as the violent local storms are of the summer. Land regions with a climate such as this have little association with the south temperate ocean zone and its comparatively equable and regular though rough climatic features. The northern land interiors can hardly be classed with the narrower oceans between them. In the presence of these regions of extremes, separated by oceans of relatively equable climate, and unrepresented in the southern hemisphere, the effort to divide the earth into symmetrical climatic zones bounded by latitude lines can hardly be accomplished.

336. The coasts of the north temperate zone. The extreme, continental or interior climates, as they are variously called, of the interior lands in the north temperate zone are strongly contrasted with the relatively equable climate of the western coastal lands in middle latitudes, particularly with that of the favored western coast of Europe, as illustrated in Fig. 18. There, on the same latitude with interiors having a January mean temperature of —10° or —20°, the coast of France and the British Isles enjoy a mean of 40° or 50°; and in summer when the interiors have July means of 70° or 80°, the favored coast rises only to 60° or 70°. Like the temperate oceans which they adjoin, their winters are wetter than their summers; but they do not suffer from periodical droughts, like the sub-tropical western coasts nearer the equator. The air is damp, with much cloudiness, especially in winter. In moderation of mean annual temperature range, the western coast of Europe, and to a somewhat less degree the western coast of North America, vie with a great part of the torrid zone; although the daily temperatures from which the monthly means are computed exhibit variations much greater on temperate coasts than in torrid latitudes, on account of the cyclonic fluctuations of the westerly winds. Inland from the open and irregular coast of western Europe, this mild climate gradually merges across the lowlands into the more severe climate of the interior; the temperatures vary over a greater range; the rainfall for a time is nearly equably distributed over the year, and then takes

on a summer maximum; the frequent cloudiness and high humidity of the coast is exchanged for clearer skies and drier air. A similar change is found in passing eastward from our Pacific coast, but here it is made abruptly. On crossing the lofty mountains which here lie so little distance inland, we enter at once on the extreme climate of the interior; the torrid heat of Arizona and the extreme cold of the Mackenzie basin with its excessive annual range lying only a few hundred miles from the mild litoral belt by the ocean.

The eastern coasts of the temperate continents, represented by our Atlantic seaboard and that of eastern Canada, might be classified with the interiors, from which they derive so many of their climatic features; but for the purpose of emphasizing their contrasts with the western coasts, they are here given a special paragraph. They partake of the strong temperature ranges, both annual and cyclonic, that characterize the interiors, because the general drift of the winds is here from land to sea. Not only so; the seasonal shifts of the winds (Sect. 155) here intensify the seasonal variations of temperature; in summer, the prevailing winds are from the over-warm southwest lands; in winter, from the over-cold northwest lands. In this, they are reversed from the relations obtaining on the western coasts; the summer winds there come prevailingly from the little-warmed northern seas; and the winter winds from the little-cooled seas of lower latitudes. The rainfall is generally well distributed through the year along our eastern coast; but on the corresponding coast of China, the winters are relatively dry, and the rains of summer are in the control of the southerly monsoon.

One of the distinct peculiarities of the climate of our eastern temperate coasts is its rapid increase of severity poleward. In spite of a strong range of temperature, our southern Atlantic coast must be included among the better favored regions of the world; but in passing northward, over a latitude range no more than that from Morocco to Scotland, we pass from Carolina to Labrador; and there, opposite to the mild climate of Great Britain, we find inhospitable shores around a forbidding interior; England, a land of mild and genial climate, in which so small a river as the Thames is only exceptionally frozen over; Labrador, an almost uninhabitable region, with a January mean but little over zero, swept over by harsh winds from a vast snow-covered interior. When this striking contrast was first recognized two hundred years ago, it so strongly impressed Halley, the eminent English astronomer of that time, that he thought the cold of northeastern America resulted from the North Pole once having occupied that part of the earth's surface.

The contrasts thus presented on an east and west line in middle temperate latitudes deserve especial attention, and again serve to illustrate the difficulty of fitting any simple system of climatic zones to the complications of nature. The latitude circle of 50° N, in summer or winter, traverses regions so dissimilar that to include them all under a single zone would defeat the object of this

chapter. Beginning on the equable Atlantic waters, the circle enters middle Europe where the history of the last thousand years has witnessed the greatest progress that the world has ever seen. It crosses the broad deserts of central Asia, where the monotonous extent of too much land, too far from the vapor-yielding seas, degrades its inhabitants, and holds them close down to barbarism. Emerging on the Pacific coast, it finds the more populous countries to the south, its own climate being too harsh for easy occupation. After crossing the broad and equable northern Pacific, it traverses the narrow belt of tempered and moist climate in British Columbia, and then beyond a rugged mountain region, with many snowy peaks, it discovers the severe interior climate of the Saskatchewan, and the sparsely settled district between our Great Lakes and Hudson Bay. On the desolate Labrador coast, the better favored climate and populous regions are again found to lie further to the south. As far as habitability is concerned, this middle temperate belt contains climatic variations almost as great as are encountered in passing from the equator to the pole.

337. The mountain climate of the temperate zone is marked by an increased intensity of insolation, a decrease of temperature and an increase, up to certain altitudes, of precipitation similar to that already described for the torrid zone. Ascending from the dry sub-tropical lowlands of southern California, one soon leaves the orange groves, passes up the slopes of the Sierra Nevada through forests at first deciduous, but in which coniferous trees rapidly increase in numbers, and at last above the tree-line, snow-fields are found on the higher peaks that last through the year from the heavy fall of winter. The lofty plateaus of Arizona and southern Utah rise above desert valleys and bear abundant forests, supported by a sufficient rainfall; one of the plateaus is well named the Aquarius. The volcanic summit of San Francisco mountain in Arizona rises to an elevation of 12,800 feet, a mile above the desert table land around its foot. Well-marked zones of vegetation, including a belt of heavy forest, have been recognized on its conical slopes; they stand obliquely, a little lower on the shady northeast than on the sunny southwest side. The uppermost of these zones, above the tree line, contains a number of Arctic plants, nine of which are identical with plants brought by General Greely from Lady Franklin Bay, near latitude 82° N.

Besides these features of reduced temperature and increased precipitation, there is an increased windiness with a nocturnal maximum, and during clear weather an active evaporation, which in part compensated for the frequently increased cloudiness. The mean diurnal and annual ranges of temperature are less than on adjacent interior lowlands. Lofty plateaus differ from mountain peaks of similar altitude, by a relatively higher temperature, an increased range of temperature and a distinct diurnal wind period, with maximum in the afternoon.

The variations of temperature in the forest air and in the soil beneath are less than in the surrounding district. In mountainous regions, the presence of forests is important in restraining floods and in holding the soil on the slopes; but in our western semi-arid plains, it is doubtful if even these indirect climatic conditions will be seriously affected by any possible tree planting; much less will the amount of rainfall be changed. The more general movements of the lower atmosphere, on which temperature and rainfall so largely depend, are practically unchanged by so slight a thing as a forest covering.

341. Periodic variations of climate. It is a popular notion that our climate is changing. The winters, for example, are often said to be less severe than when old men were boys; or the Gulf Stream is thought to shift its course and thereby affect the climate on our eastern coast. These errors arise in the first place from the natural exaggeration of past events, and from the disposition to forget facts of ordinary value and dwell on exceptional occurrences; and in the second place from a certain credulity regarding unseen and remote processes. While it is well known that the course of the Gulf Stream varies by small amounts and for short periods, it is also well known that its average course depends on long-lived controls, such as the shape of the ocean basin and the strength of the general winds; the latter in turn depends on sunshine, and there is no reason to think that either the ocean basins or the strength of sunshine fluctuate to the extent implied in popular beliefs. Records of rainfall and temperature maintained for the longest series of years do not confirm the common ideas regarding our winters. The averages for decades in the early part of the century are essentially equal to those now obtained. If slight differences appear, it is much more likely that they are due to changes in the instruments used, or in their surroundings, as by the growth of trees, or the building of houses, or to changes in the residence of the observers, than that they are due to actual changes in terrestrial or non-terrestrial controls of climate. It is true that slight fluctuations of rainfall and temperature in nearly eleven years, corresponding to the sun-spot cycle, have been made out at certain stations for a moderate number of periods; but the fluctuations have not yet been shown to be general, uniform, and persistent. A longer variation is indicated over Europe and in certain other countries in a period of thirty-six or thirty-seven years, as shown by Brückner's review of all available records of dry and wet years, high and low stages in rivers, abundant and scanty crops, etc.; but at least another century will be needed fully to confirm this result and to extend it over the world. The middle dates of Brückner's periods of slightly greater rainfall and lower temperature are 1671–75, 1696–1700, 1741–45, 1766–70, 1816–20, 1851–55, 1880; and of less rainfall and higher temperature, 1681–85, 1726–30, 1756–60, 1786–90, 1820–30, 1861–65. One of the most

prevailing low temperature; the sandy and stony deserts of trade wind belts, of areas in the lee of lofty mountain ranges and of interior basins, depend on the prevailing low humidity. The occasional salt deserts of the world also belong in the latter class. In milder and moister climates, all these areas might support a plentiful fauna and flora.

The first essential condition for the support of land plants and after them of animals, is the preparation of a covering of soil, formed by the disintegration of the underlying rock; and there are no known rocks that will not in a sufficiently long time weather into a loose plant-supporting soil. Locally, the loosened material may be carried away as fast as it is formed, as from the rocky slopes of mountains and from the ledges of hills; but this is exceptional. The second condition is a fitting climate. In too dry a region, the soil may be produced, but it lies sterile; or the finer parts may be blown away, leaving a sandy or stony surface, nearly or quite barren. In too cold a climate, the ground is frozen, snow accumulates over it, and nearly all life is excluded. But where the rainfall is sufficient, — that is, over 10 or 15 inches a year, — and where the cold is not excessive, so that during at least a part of the year the ground is melted, and the temperature is sufficiently high to encourage flowering and fruiting, plants and animals may survive, their forms and habits being specially adapted to the unfavorable conditions in which they live.

From these limiting conditions, an increasing variety of life is found with increasing warmth and humidity, until the luxuriance of the sub-equatorial belt is reached. It is noteworthy that a moderately warm summer alternating with a severe winter, such as occurs in the northern continental interiors, is much more favorable for the development of varied forms of life than the equable but inhospitable climate of the remote islands in the boundless seas of the south temperate zone. It is also instructive to see that the climatic conditions most favorable for man are not those of the equatorial lands, where the temperature is enervating and where the support of life presents no difficulty and calls for little forethought; but those of the northern temperate zone, where the rigors of the coming winter season call for a thoughtful preparation during the preceding summer.

340. Local control of climate. In the chapter on rainfall, mention was made of the attempts to produce artificial rain; either temporarily, as by extensive fires or by explosions, or permanently, by planting trees. Observations thus far made do not give encouragement to these projects. On the other hand, it has been proved that areas of forest amid an open country are in a small way conservative in their influence, decreasing the suddenness of the changes that take place about them. The rain that falls does not run away so quickly, and therefore not only provides a better supply for the steady running of streams, but also decreases the loss of soil by surface washing.

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interesting indications of this thirty-six year period is found in the variation in the lengths of Swiss glaciers; the periods of extension being 1760–86, 1811–22, 1840–55, 1880–; and the periods of retreat being 1750–67, 1800–12, 1822–44, 1855–80. It is found that the shorter glaciers are the first to feel the change in their upper snow supply, and to lengthen or shorten accordingly; hence all the glaciers of the Alps are not retreating and advancing at the same time. Yet for a few years, the longest and the shortest advance or retreat together; thus from 1815 to 1818, all the Alpine glaciers were advancing; from 1822 to 1825, all were retreating; from 1848 to 1850, all were advancing again; about 1875, all were retreating; and now another general advance is approaching.

342. Secular variation of climate. Ancient historic records around the Mediterranean Sea have been accepted by special students of this question as indicating a general decrease of rainfall there in the last three thousand years. In northern Africa, the remains of cities imply a greater population than can now exist in that desert region; ruins of aqueducts and irrigating canals are found in districts where there are now no sufficient streams to supply them; ancient records mention the presence of certain animals there, from which a less arid climate than the present would be inferred. Some would ascribe this climatic change to the ancient destruction of forests; but there is no direct evidence of the existence or the destruction of forests along the northern coasts of Africa; if they once grew there and were destroyed by man, it is quite as reasonable to suppose that they were then dwindling away and could not naturally restore their growth under an increasingly unfavorable climate, as to believe that the change of climate was entirely due to their destruction.

343. Geological changes of climate. On passing from ancient historic records back to recent geological records, abundant evidence is obtained of climatic changes. In pleistocene time, that division of the geological scale next preceding the present, the northwestern part of Europe, the northeastern part of North America, and certain other regions of less extent, were covered with ice, much as Greenland is to-day; and many interior basins where an arid climate now prevails were then flooded with broad lakes. This seems to have happened not only once, but repeatedly; the records appearing to be more complicated the more closely they are studied. While it is not certain that the lacustrine conditions of the interior basins were coincident with the glaciation of North America and Europe, it is highly probable that such was the case; and that the climate then prevailing was somewhat colder than now; thus increasing the length of the cold season when snow might accumulate, decreasing the season when it would be melted, and diminishing the ratio of evaporation to rainfall in the now arid basins.

The causes of the pleistocene climatic variations have been sought in a change in the altitude of land masses with respect to sea level, in a change in the sun's heat, in a change in the form of the earth's orbit, and in various other conditions; but no general agreement is yet reached regarding them. The changes in the altitude of the land would manifestly induce an increase of snowfall; but it does not appear from other evidence that the glaciated regions were higher when the ice accumulated on them than they are now. The variation in the form of the earth's orbit is regarded by many students as the most effective control of pleistocene climatic changes, and if such is the fact, similar changes should have been of repeated occurrence in the geological past. The variations referred to are an increase in the eccentricity of the orbit, whereby the inequality of the periods between the equinoxes and the inequality of insolation at perihelion and aphelion (Sect. 27) may be increased; and a slow change in the date of the equinoxes, whereby first one hemisphere and then the other would have its winter in aphelion. At present, the eccentricity is moderate; and the northern or land hemisphere has its winter in perihelion. If at a time of great eccentricity, the northern hemisphere should have its winter in aphelion, it is possible that the accumulation of snow and ice during so long and severe a season might not be consumed in the relatively short but hot summer; thus snow and ice would accumulate, until astronomical changes brought back milder conditions. All this subject is one that still needs extended investigation.

In more remote geological ages, various climatic changes are indicated. Coal beds were formed in Greenland; forests grew in the deserts of Arizona and Egypt, ice-borne boulders were carried into the sea in which the chalk of England was formed. These and similar changes are presumably to be explained in great part by changes in the distribution and form of the land areas, and consequent changes in the course of the winds and the ocean currents; as well as by astronomical changes of the kind just mentioned.

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